Volume 10, Number 6
November–December 2002

ISSN: 0869-5938

STRATIGRAPHY AND GEOLOGICAL CORRELATION

English Translation of *Stratigrafiya. Geologicheskaya Korrelyatsiya*

Editor-in-Chief
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http://www.maik.ru

Journal of the Fundamental and Applied Aspects of Stratigraphy and the Correlation of Geologic Events and Processes in Time and Space

МАИК "НАУКА/INTERPERIODICA" PUBLISHING

Stratigraphy and Geological Correlation

(Stratigrafiya. Geologicheskaya Korrelyatsiya)

ISSN: 0869-5938

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Stratigraphy and Geological Correlation is abstracted and/or indexed in *Petroleum Abstracts*, *Research Alert*, *SciSearch*, *Current Contents* (*Physical, Chemical*, and *Earth Sciences*) and *World Translations Index*.

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PUBLISHER
INTERNATIONAL ACADEMIC PUBLISHING
COMPANY "NAUKA/INTERPERIODICA"
PROFSOYUZNAYA UL. 90, MOSCOW, 117997 RUSSIA
PHONE: 7 (095) 336-1600; FAX: 7 (095) 336-0666
E-mail: compmg@maik.ru
Web Site: http://www.maik.ru

SUBSCRIPTION PRICES (2002, Volume 10, 6 issues)
US $1165.00 (North America), $1335.00 (elsewhere). Subscriptions may be entered at any time. Orders may be placed at agencies or directly. Cancellations not allowed except for duplicate payment. Prepayment required. Advance payment required in US$/US bank. MasterCard/Visa accepted. Make checks payable to *Stratigraphy and Geological Correlation* or Interperiodica.

Stratigraphy and Geological Correlation (ISSN: 0869-5938) is published bimonthly for US $1165.00 (North America), outside North America for $1335.00 per year by MAIK "Nauka/Interperiodica", 5724 Hwy. 280 East, Birmingham, Alabama 35242-5724 USA. Periodical postage paid at Birmingham, Alabama, and additional mailing offices. POSTMASTER: Send address changes to *Stratigraphy and Geological Correlation*, P.O. Box 1831, Birmingham, Alabama 35201 USA.

The journal was founded in 1993.

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Printed in the USA by Pleiades Publishing, Inc.

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Different Approaches to Perfection of the General Stratigraphic Scale

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Received June 19, 2001

Abstract—Different approaches (American and European) to perfection of the General stratigraphic scale (GSS) are discussed. Practice of the International Stratigraphic Commission that consists, according to recommendations of "International guides," in "hammering in the golden spikes" (GSSP) in rock sequences of monotonous lithology at the levels marked by changes in composition of fossil species (mainly conodonts), which are thought to be evolutionary and independent of the environmental influence, is criticized. Boundaries between series of the Devonian System and between the Permian and Triassic systems are considered as examples, and it is shown that such an approach to recognition of boundaries between major stratigraphic subdivisions violates the priority principle and eventually leads to destabilization and complete reorganization of the GSS. Principles used to specify the GSS are not only of the abstract theoretical interest, but they directly influence practice of geological mapping and other geological investigations.

Key words: General stratigraphic scale, correlation, Global boundary stratotype section and point, Devonian System, Permian System, Triassic System.

INTRODUCTION

The Russian Stratigraphic Code determines the General stratigraphic scale as "an integrity of general stratigraphic subdivisions (in their complete volumes without gaps and overlapping) arranged in their stratigraphic succession and hierarchical subordination" (*Stratigraficheskii kodeks...*, 1992, p. 22). It is believed that general stratigraphic subdivisions are potentially of planetary significance and can be used by geologists from different countries. Therefore, the scale composed of these units is also termed as the International stratigraphic scale (ISS). International stratigraphic and corresponding geochronological scales were elaborated in general more than a century ago and were approved at the Second session of the International Geological Congress in Bologna, Italy in 1981. Since that time, the International stratigraphic scale served as a basis for any geological researches. In opinion of Leonov (1973), it did not experience any substantial changes until the 1970s. All introduced changes and additions were insignificant and did not violate the general principle the scale is based on.

The second half of the 20th century and particularly the last decades were characterized by intense researches aimed at perfection of the general stratigraphic scale. The activity of the International Stratigraphic Commission whose purpose was to specify boundaries between geological systems, series, and stages also intensified. Many Soviet geologists were involved in different international projects. Simultaneously, investigations of precisely this period showed differences in approaches to solution of main stratigraphic problems. Two different concepts that appeared were conditionally termed the "American" and "European" ones (Zhamoida and Menner, 1974).

The "American" concept, as it is reflected in the American Stratigraphic Code and in the International Stratigraphic Guide (1976) translated into Russian under the name *Mezhdunarodnyi stratigraficheskii spravochnik* (1978), is based on the idea of multiple independent stratigraphic categories of different significance (lithostratigraphic, biostratigraphic, chronostratigraphic). It also assumes the artificial character of boundaries between units of the general stratigraphic (geochronological) scale, which should be distinguished in a continuous succession of preferably marine sediments. The priority principle obligatory for the scientific terminology is virtually ignored. Serious changes have not been introduced into the second edition of the guide (*International Stratigraphic...*, 1994; Zhamoida, 1996).

The "European" concept that is reflected in stratigraphic codes of the USSR and Russia (*Stratigraficheskii kodeks...*, 1977, 1992) considers stratigraphy as a single fundamental branch of geological sciences, but not as many *stratigraphies* independent from each other and based on different research methods. According to this concept, main stratigraphic subdivisions, including global, regional, and local ones, are of a geosystem nature, because they reflect certain stages in development of the geosphere as a whole and of its separate parts. Special stratigraphic units are estab-

lished using special methods and are auxiliary to main subdivisions.

Many researchers (Zhamoida and Menner, 1974; Menner, 1978; Zhamoida and Moiseeva, 1980; Sokolov *et al.*, 1992) expressed a hope for the gradual convergence of the "American" and "European" concepts and elaboration of a common approach to main stratigraphic problems. The further events showed, however, that proposals of many European (including Russian) stratigraphers are ignored. Members of the Working group on editing of the International Stratigraphic Guide (including representatives of Australia and the USSR) who disagreed with views of its leaders lost opportunity to participate in this work, and the published guide reflects only one standpoint.

The further development of principles for selecting global stratotypes of boundaries between ISS units, which are formulated in the *International Stratigraphic Guide*, can be seen in *Guidelines for establishment of global chronostratigraphic standards* (Remane *et al.*, 1996). This brief *manifest of contemporary chronostratigraphy* contains some rightful provisions. It states that the priority should not be ignored completely, when selecting the boundary stratotypes,[1] and that other criteria, e.g., levels of particular events and others, should be used along with paleontological ones. These good intentions however disappear from solutions under the pressure of *interests* of the voting majority, but not of *scientific arguments*. It is absolutely clear that voting is not a scientific method of solving problems. Attempts to solve scientific problems by voting are based on the assumed artificial character of stratigraphic boundaries and *convenience principle*. According to Leonov (1973, p. 517), "the personal convenience meant here is usually inconvenient for others. It is apparent that common agreed principles cannot be elaborated for the international scale based on such approach."

Main drawbacks of mentioned guidelines consist in ignoring the European concept of stratigraphic classification that is based on a tremendous experience of European (including Russian) geologists and resolutions of international geological congresses, besides the neglecting of priority principle. All this played a negative role during elaboration of a new variant of the general stratigraphic (geochronological) scale that was presented at the last IGC session. Many points of this project appeared to be inappropriate for Russian geologists, as it can be partly seen from the Interdepartmental Stratigraphic Committee resolution adopted in 1999.

We do not intend in this paper to discuss once again the problem of the artificial character of boundaries between ISS units and persuade our opponents in fallibility of their views. This is impossible in view of the character of contemporary discussions. The aim of the paper is to demonstrate negative consequences that result from the modern practice to approve new boundaries between geological systems and their series based on recommendations of "international guides." We want as well to discuss the topical question: how should stratigraphers who are categorically against some novelties behave in order to protect the domestic stratigraphy from harm?

Any researcher who realizes the dialectics of scientific knowledge development should understand that extreme approaches could never serve as a basis for constructive solutions. Nobody amid adherents of the "natural" character of the ISS subdivisions do not consider the latter as corresponding to similar one-rank stages in development of lithosphere and biosphere. Otherwise, attempts (still unsuccessful) of radical reorganization of the geochronological scale based on major tectonic, biological, and other events would be less numerous. The contemporary geological science is not ready for such reorganization. Simultaneously, irregular patterns (and certain periodicity) of earth crust, hydrosphere, and atmosphere evolution represent the empirically established regularity.

Elaboration of the global stratigraphic scale developed spontaneously. Views on position of boundaries between stratigraphic subdivisions changed many times. Significance of events that substantiate boundaries between systems and series is far from being unambiguous. Moreover, boundaries between some series are undoubtedly more important in the historical–geological respect than boundaries between systems. These and many other facts allow the International stratigraphic scale to be considered as representing a *conditional* construction. Simultaneously, this conclusion has nothing to do with proposition to consider the geochronological scale as the *artificial* construction and with attempts to completely exclude the principle of "natural character" peculiar of stratigraphic subdivisions from stratigraphy, as it was done in the American and international stratigraphic codes. Considering the global stratigraphic scale as a conditional construction, G.P. Leonov simultaneously emphasized that "more or less complete correspondence of most subdivisions in the international scale to certain stages in geological development of separate regions is evident from the entire history of defining of these units, being out of doubts" (Leonov, 1973, p. 453).

In his fundamental work on stratigraphy, Leonov managed to avoid extremities, "walked along the razor blade," and formulated principles of converting the International stratigraphic scale into the universal, stable, and sufficiently detailed system of geochronological classification. He suggested "(1) to consider the geochronological scale as a conditional construction, (2) to follow the regional–stratigraphic principle of determining the range and boundaries of its units based

[1] However, when determining ranges of ISS units, the authors likely mean preferential variants of new global boundary stratotypes rather than the priority principle or "historical method" that was certified in resolutions of the 8th International Geological Congress and in all classic works.

on relevant stratigraphic standards corresponding to particular regional stratigraphic units, and (3) to keep stable the ranges and boundary positions in the international scale based on the priority principle" (Leonov, 1973, p 526). If international geological organizations were following these recommendations, situation with the ISS would be different. Unfortunately, the International Stratigraphic Commission took another way, and as a result we have now what we have.

In order to retain the traditional regional–stratigraphic principle of establishing the unit boundaries, laborious researches are needed. One should pass, using correlation based on the principle of chronological interchangeability of certain indications (Meyen, 1989), from sections with hiatuses and sharp facies variations, which were used to establish the units originally, to other sections, where hiatuses are minimal or (in the ideal case) absent at all, and where indications of corresponding events are expressed in a different form. Precisely such sections are appropriate to select the GSSP.

Nevertheless, the International Stratigraphic Commission and its subcommissions recommend to "hammer in the golden spikes" in the parts of sections of uniform lithology, where successions of fossil species are allegedly independent on the environmental influence, being exclusively evolutionary. Such a way of searching for the "ideal" boundaries is widely used, although it strongly resembles an old anecdote about a person who searched for the purse under the lantern, because there is lighter, but not in the place of loss. Rock members, which fit the above requirement, are rare near traditional boundaries between systems and series, being more frequent in their middle parts. Accordingly, the newly selected levels appear to be substantially displaced relative to the former. An approval of such levels as corresponding to boundaries between major stratigraphic subdivisions results in a brute violation of the priority principle and, eventually, in destabilization and complete reorganization of the general stratigraphic scale (Karaulov, 1994).

Refusing to use the historical–geological method of establishing boundaries between major stratigraphic subdivisions of the ISS, we objectively meet the situation, when the number of features, which can potentially characterize a boundary, radically decreases. For the sake of the truth, it should be noted that the last achievements in development of non-paleontological methods, for instance, the Ir geochemical anomaly established across the Cretaceous–Paleogene boundary, a sharp decrease of $\delta^{13}C$ values across the Permian–Triassic boundary, or recognition of narrow paleomagnetic zones, clearly demonstrate their significance for precising the boundary levels (Remane *et al.*, 1996). Nevertheless, the leading role still belongs, as before, to paleontological criteria. The latter are often understood as the *first appearance* of a species selected arbitrarily (by agreement or voting) that actually means *the first occurrence level* of that species in a particular (stratotype) section. In the majority of sections, barren layers and members separate sediments bearing paleontological remains. To avoid the problem to what adjacent fossiliferous units such intervals should be referred, many researchers prefer to select the boundary stratotypes in condensed sections, which are extremely rare, having very low correlation potential based on fossil remains of some "ultrastratigraphic" groups only.

Conodonts, which were under keen attention in last decades, often to the detriment of other similarly important fossils, represent one of such fossil groups.[2] Conodont remains are found to characterize previously "barren" intervals in different marine sections composed of terrigenous, calcareous, and siliceous sediments—the fact of a positive significance for stratigraphy. On the other hand, the researchers who study conodonts consider them a priori as representing the *archistratigraphic* group and attribute a supreme role to the conodont zonal standards. Nevertheless, drawbacks of the stage stratigraphic scale of the Permian System that is based exclusively on conodonts and formal successive replacement of their species in a single phylogenetic lineage were recently discussed by Leven (2001). Traditional fossils such as ammonites, which served successfully before and serve now as a basis for subdivision of Paleozoic and, particularly, Cenozoic sequences, are ignored in this case.

As a result, practical geologists are deprived from opportunity to trace confidently boundaries of the ISS units and to show them in geological maps. Paleontological methods are undoubtedly decisive for the Phanerozoic stratigraphy. Nevertheless, if we really want to establish boundaries between major stratigraphic subdivisions of the global nature, we should take into consideration stages in development of the whole organic world, verifying our conclusions by other, first of all, lithological–formational methods, which allow the different-rank historical–geological stages to be objectively revealed.

There are many examples when methods adopted by international organizations for defining boundaries between major ISS units negatively influence practical geological investigations. We will consider here only some of them concerning those intervals of the stratigraphic scale, which we are better acquainted with from our experience. Before the pertinent discussion, let us consider the most general problem of positioning the Precambrian–Phanerozoic boundary.

[2] Conodonts are the multielement remains of circumoral apparatuses of ancient animals whose integral biological system, but not its disintegrated elements, is now under incipient deciphering only (e.g. Orchard and Rieber, 1998). Stratigraphically most informative are their paired elements Pa. At present, there is a tendency to define two types of conodont zonal scales based on pelagic and shallow-water forms.

ON THE CAMBRIAN LOWER BOUNDARY

The most important problem related to specification of boundaries between geological systems has always been and remains now the problem of the lower boundary of the Cambrian System (and of the entire Phanerozoic as well). We are not experts in this field and just can discuss some aspects concerning the method applied for the boundary establishing. The All-Union meeting held in Ufa in 1967 decided that, defining the lower boundary of the Cambrian System, geologists should take into consideration paleontological data first of all, regarding the appearance of skeletal fauna as the most important criterion. Similarly to members of the International Working Group on the Precambrian–Cambrian boundary, participants of that meeting considered it unnecessary to use the historical–geological criteria (Rozanov *et al.*, 1976). Searches for the global stratotype of this boundary lasted more than 30 years and ended with decision to select the necessary point in the Newfoundland section and to define it based on changes in ichnofossil assemblages (fossil traces of organisms of unclear taxonomic affinity); this appeared to be absolutely unacceptable for the Russian and many other stratigraphers (Rozanov *et al.*, 1997).

This could be expected, because the decision was made by voting under influence of researchers occupying key position in relevant international organizations who had their particular, not always scientific and purposeful interests. Discussing the problem of the boundary in question, Sokolov (1997, p. 5) wrote: "From my standpoint, it is defined now as a standard one quite accidentally at the base of the *Phycodes pedum* Zone of Newfoundland, thus being based on fossilized traces of unknown organism of a wide stratigraphic range. Of course, this boundary stratotype should be revised with due account for traditional biostratigraphic requirements, though the decision that *physical zone at the base of the Cambrian System should be reliably recognizable worldwide remains principally important.*"

From our point of view, the refusal from the historical–geological (regional-, event-, and eco-stratigraphic) criteria was premature and erroneous, because precisely the mass appearance of skeletal organisms at the base of the Tommotian Stage (Rozanov *et al.*, 1969) is the most important argument in favor of placing the considered boundary at this level from both the paleontological and historical–geological standpoints. From the last standpoint, a hiatus present at the base of most Tommotian sections only emphasizes the boundary significance and does not represent an obstacle for its approval as a lower boundary of the Phanerozoic, especially if we take into consideration the fact that there are sections where this hiatus is minimal or absent at all. In any case, only relevant resolutions of the Interdepartmental Stratigraphic Committee are appropriate for geologists who carry out investigations in the Siberian platform and other regions of Russia.

BOUNDARIES OF THE DEVONIAN SYSTEM AND ITS SERIES

The problem of the Devonian lower boundary has a long-lasting and very pictorial history considered in detail by Leonov (1973). A special committee of the IUGS Stratigraphic Commission that was organized in 1960 had to solve the problem of boundary separating the Silurian and Devonian systems. Since that time, several national and international programs of stratigraphic research across the Silurian–Devonian boundary deposits, which searched for a satisfactory solution of the boundary problem in continuous marine sequences, were accomplished in different countries (*Granitsa silura...*, 1971).

These investigations yielded new extensive materials, according to which the Silurian–Devonian boundary was placed at the first appearance level of *Monograptus uniformis*, the zonal graptolite form, between the Pridolian and Lochkovian stages of the Pragian syncline. This decision was approved at the Montreal session of the International Geological Congress in 1972. The GSSP of the boundary was established in the Klonk section 35 km southwest of Prague. It should be emphasized that in this case the boundary between stages, which were previously defined as formations, was selected to represent a boundary between major stratigraphic subdivisions of the ISS, i.e., the regional–stratigraphic principle has not been violated.

The selected boundary was accepted by the majority of stratigraphers. It is close to the traditional one (slightly higher than that accepted in England, slightly lower as compared with the level previously adopted in the USSR, and approximately corresponding the Belgian–German variant) and bears a certain historical–geological meaning (coincides with the beginning of one of the transgressive–regressive cycles and with a substantial biota renewal), being widely traceable based on several features, although significance of corresponding events is noticeably lower in comparison with some other Devonian boundary events.

Resolutions on series boundaries of the Devonian System, which were adopted by the International Subcommission of the Devonian Stratigraphy (SDS) organized in 1972, are more controversial. When selecting these boundaries, the subcommission neglected the regional–stratigraphic approach and proposed several occurrence levels of conodont species and subspecies as potential boundary stratotypes, though they do not correspond in most cases to significant historical–geological and biotic events and, consequently, have a low correlation potential.

Originally, the SDS proposed four levels, which may represent the lower boundary of the Middle Devonian: (1) the base of the *costatus costatus* conodont zone; (2) the base of the upper subzone of the *costatus patulus* Zone; (3) the base of the *costatus patulus* Zone; and (4) the first occurrence level of *Polygnathus dehiscens* and *Monograptus yukonensis*. In 1979, only two

variants were left for selecting the Lower–Middle Devonian boundary: (1) the base of the *costatus patulus* Zone corresponding to the base of the Couvinian Stage in Ardennes; and (2) the base of the *costatus partitus* presumably corresponding to the base of the Lauch Beds in the Eifel Hills (Rzonsnitskaya, 1983). In the same year, resolution was voted to place the lower boundary of the Eifelian Stage and, correspondingly, of the Middle Devonian, at the first occurrence level of *Polygnathus costatus partitus*. This level is recorded 1.9 m below the lower boundary of the Lauch Beds in the Wetteldorf section of the Eifelian Mountains that represents a stratotype of the Eifelian Stage (Ziegler and Klapper, 1985).

Discussion of the problem of the Middle Devonian lower boundary showed, on the one hand, the invalidity of the extended *Eifelian* Stage adopted in the USSR and gave birth, on the other, to the problem of the lower stage of the Middle Devonian (*Nizhnii yarus...*, 1983) that is still waiting its solution. The problem is rooted in the fact that the boundary characterizing distinct changes in development of many groups of fossil organisms and almost a universal change of geological formations was considered in the USSR as a boundary between the lower and middle series of the Devonian System. In its new understanding, it turned out to be significantly below the base of the Eifelian Stage, and the base of the Eifelian Stage and Middle Devonian appeared to be indistinguishable and untraceable in the territory of Russia and adjacent regions of North Eurasia based on paleontological or historical–geological criteria.

At the same time, there are distinct boundaries related to paleogeographic and paleobiological events, which are easily traceable not only in the European stratotype sections, but worldwide as well. Johnson *et al.* (1985) compiled an eustatic curve that reflects alternation of major Devonian transgression and regressions in North America and West Europe. Similar curve was obtained by House (1985) when he studied evolutionary stages of Paleozoic ammonoids. Studying independently the Devonian history of the Uralian–Mongolian belt and adjacent regions of North Eurasia, one of us established several phases of considerable extension and reduction of Devonian sea basins related to eustatic sea-level fluctuations, which have been induced by tectonics (Karaulov, 1988; Karaulov and Gretschishnikova, 1997). Boundaries between largest transgressive–regressive cycles in the curves compiled for different continents coincide that implies their planetary character.

The International Symposium "Eustatic sea-level fluctuations during the Devonian" held in 1994 (*Sbornik tezisov...*, 1994) confirmed the validity of above conclusion. Consequently, the statement that distinct boundaries between stratigraphic units of different sections are controlled by local factors and cannot be traced in other regions is only partly true. Most significant historical–geological boundaries are related to global events and traceable almost worldwide. It should be kept in mind, however, that indications of these events can very differ in different-type sections (Karaulov, 1977).

Returning to the problem of the lower stage of the Middle Devonian, let us recollect that serious arguments in favor of placing the lower boundary of Middle Devonian at the base of either Couvinian (Rzhonsnitskaya, 1983) or Dalejian (Bogoslovskii, 1983) stages were proposed long ago at the initial stage of the discussion. This level is very close to the first of most important historical–geological boundaries mentioned above and to the traditional Lower–Middle Devonian boundary accepted in Russia. It is also not incidental that the Couvinian Stage that included then the Ierge Beds of graywackes was placed at the base of the Middle Devonian already in the first Devonian stratigraphic scales (Leonov, 1973, p. 268).

The subsequent stratigraphic practice showed, as it could be expected, the vainness of attempts to fix the lower boundary of the Middle Devonian in Russia at the level recommended by the SDS. According to the resolution of the Interdepartmental Stratigraphic Meeting dedicated to the Middle and Upper Paleozoic deposits of the Russian platform, the base of the Middle Devonian is placed at the base of the Kojva Horizon of the upper Emsian corresponding to the base of the *Zdimir pseudobaschkiricus–Megastrophia uralensis* brachiopod zone or to that of the standard *patulus* conodont zone (*Reshenie mezhvedomstvennogo...*, 1990). It is absolutely clear that all attempts to place the Lower–Middle Devonian boundary closer to the level approved by the SDS are futile and result in destabilization only. The boundary displacement from traditional level to the base of the *Zdimir pseudobaschkiricus–Megastrophia uralensis* Zone is unable to solve the problem as well, because the latter is situated one standard zone lower than the boundary recommended by the SDS, being far from the convenience of its tracing (Karaulov, 1994).

So, where is an exit from the situation? It is evidently unreasonable to go back to the problem of the Eifelian Stage lower boundary. Taking into consideration an appeal for opinions about rational division of some stages into substages (the SDS Newsletter no. 13) and obvious great significance of historical–geological criteria indicative of boundary events between stratigraphic units, we may suggest to subdivide the Emsian Stage, the lower boundary of which has not been defined so far, into the Zlichovian and Dalejian stages of the Czech stratigraphic scale. The SDS already uses Czech stage units in the general scale. The lower boundary of the Dalejian Stage corresponds to the base of the *Anarcestes* goniatite zone, is close to the base of standard *inversus* zone in the conodont zonation, and coincides with the boundary between distinct historical–geological stages. Precisely this level used to be

regarded in the former USSR as corresponding to the lower boundary of the Middle Devonian.

Similarly intricate situation concerns the Middle–Upper Devonian boundary. Until now, this boundary in Russia was reliably placed at the base of the Pashyja Horizon and its age analogues (*Stratigrafiya SSSR...*, 1973) and correlated with the base of the Fromelen Beds, i.e., with the traditional boundary between the Givetian and Frasnian stages. The situation changed when the SDS decided to place the lower boundary of the Frasnian Stage and Upper Devonian at a considerably higher level (at the base of the standard *L. asymmetricus* conodont zone or in the middle of the *falsiovalis* Zone).

When selecting the Middle–Upper Devonian boundary, the SDS used the formal criterion—the first occurrence level of some conodont species. By voting in 1982 (seven pro and three contra), it was decided to draw this boundary at the base of the *L. asymmetricus* Zone, where the first occurrence level of *Ancyrodella rotundiloba* is recorded (Ziegler and Klapper, 1985). The boundary stratotype problem was considered to be of a second priority in this case and its solution was postponed. Subsequent samplings across the selected boundary resulted in identifications of slightly different conodont assemblages suggesting different levels for the boundary, and there appeared a necessity to revise the zonal scale (Bultynck, 1986). Somewhat later, the GSSP was selected in the Montagne Noire section (southern France). The boundary point in this section is 50 cm below the base of the *L. asymmetricus* Zone, and authors of the relevant publication considered the situation at hands as exemplifying discrepancy between intentions and practice of elaborating the conodont biostratigraphic zonation (Sandberg *et al.*, 1988). When defining "standard zones" of the Upper Devonian afterward, specialists on conodonts concentrated attention on the *Palmatolepis* lineage and placed the Upper Devonian lower boundary inside the *falsiovalis* Zone.

Specialists on conodonts probably persuaded each other that they have selected at last the ideal Middle–Upper Devonian boundary. nevertheless, they can hardly persuade practical geologists that the boundary established in one–two unique successions is precisely a universal traceable level that separates the middle and upper series of the Devonian System. While the base of the *L. asymmetricus* Zone is more or less close to the base of the Sargaevo Horizon of the East-European platform and western Urals, the new "ideal boundary" appears to be inside the Kynov and Timan horizons and can be traced nowhere. One can hardly imagine another, more vivid example of the fallaciousness of the method accepted by the SDS and ISC for "specification" of boundaries between the major ISS units.

The raised problem of the Middle–Upper Devonian boundary compelled to carry out the comparative historical–geological analysis of structure of the upper Givetian–Frasnian sections in Eurasia and other continents (Karaulov, 1989a, 1989b). The results obtained showed that several approximately isochronous levels corresponding to the initial phases of extension and deepening of Devonian basins can be outlined and widely traced. Most significant among them is one level that coincides with the traditional Givetian–Frasnian boundary. Precisely this level occurring in the middle of the *varcus* Zone characterizes a sharp change in development of ammonoids, other faunal groups, and terrestrial flora. House (1985) who specially studied biotic events in evolution of Devonian ammonoids showed that the level in question marks the *Taghanic Event.* As mentioned above, this level coinciding with the traditional Middle–Upper Devonian boundary corresponds to one of two most significant falls in the sea-level curve (Johnson *et al.*, 1985; Karaulov and Gretschishnikova, 1997).

The Middle–Upper Devonian boundary, which occupied during many decades the base of the Fromelen Beds of the Frasnian Stage or the base of the Adorphian Stage, defines the first occurrence level of the ammonoid genus *Pharciceras* and archaeopterid flora. It also marks evolutionary changes in other groups of fossil organisms, coincides with the major historical–geological event, and should be kept at the traditional level. The SDS recommendation to displace this boundary, one of the most distinct and widely traceable traditional boundaries, represents a dangerous precedent of the priority principle violation and leads to destabilization of the stratigraphic nomenclature.

The idea to separate the upper part of the "new" Givetian Stage as the autonomous Fromelen Stage with subsequent probable recurring of the latter into the Upper Devonian can represent a solution of the problem (in line with the above-mentioned Newsletter no. 13). The Fromelen Stage includes the Fromelen Formation of Ardennes, Tailfer and de Roux formations of the northern flank of the Dinant synclinorium, Pashyija and Timan horizons of the East European platform, Altchedat and Izyly horizons of the Altai–Sayan folded region, and Saryn Horizon of the Omulevskii uplift (northeastern Russia). If these proposals will be rejected by the SDS, the Interdepartmental Stratigraphic Committee of Russia should independently make a decision concerning the boundaries of the Middle Devonian Series in this country.

The Devonian–Carboniferous boundary remained debatable for a long time. At present, it is placed at the base of the standard *sulcata* conodont zone that approximately coincides with the boundary between the *Wocklumeria* and *Gattendorphia* genus-zones. A substantially higher position of this boundary, as compared with that earlier accepted in the USSR, is pregnant with certain difficulties and raises a problem of separating the upper part of the "new" wider-range Famennian Stage as an autonomous subdivision. One of the possible names for the latter is the "Strunian Stage," the lower boundary of which coincides with basal levels of

the goniatite *Climenia* and ostracod *hemisphaerica–triceratina* zones, being simultaneously close to the base of the *expansa* zone in conodont zonation. From the historical–geological viewpoint, this boundary corresponds to the initial stage of a great transgression. The higher part of the Famennian Stage that was previously referred to the basal Carboniferous is correlative with the following regional subdivisions: upper *Climenia* limestones and shales of Thuringia; Schistes de l'Epinette and Calcaire d'Etroeungt of the Dinant basin; Plava, Ozerki, Khovanskii, and Ziganskii horizons of the East European platform; basal and brachiopod beds of the Tarkhanka Formation of the Rudnyi Altai; and with other synchronous sequences (Karaulov and Gretschishnikova, 1988).

A LAME ATTEMPT TO DEFINE THE PERMIAN–TRIASSIC GSSP

As is known, the Permian–Triassic boundary layers were established in continental successions. Von Alberti (1834) began the Triassic in West Europe with the Buntsandstein (red beds) overlying the saliferous Zechstein, the upper subdivision of the "Peneen" System distinguished by Omalius d'Halloy in 1831. Seven years later, Murchison (1841) who defined the Permian System in Russia included in the latter, although with some doubts, continental red beds of the central Moscow syneclise, which have been later identified by Nikitin (1887) as the Tatarian Stage. For a long time, these deposits were considered as transitional between the Permian and Triassic Systems (Permotrias). Subsequently, it was shown that the lower part of the Nikitin's Tatarian Stage is of the Late Permian age, and now this name is retained for this interval, whereas the upper part (Vetluga and Yarensk groups, according to the contemporary nomenclature) turned out to be correlative with the Buntsandstein of the Lower Triassic (Lozovsky, 1997).

Later on, when marine standards of the Triassic were distinguished in the Tethyan realm, the base of the ammonite *Otoceras woodwardi* Zone defined in Himalayas was selected for the lower boundary of the system (Mojsisovich *et al.*, 1895). In the Boreal realm, the *Otoceras* fauna, which became known later, is represented by other species (*O. concavum* and *O. boreale*). The succession of boreal ammonites is divided in two zones, the first of which begins the Griesbachian Stage of the Lower Triassic in Canada (Tozer, 1967). In northeastern Russia, the lower *Otoceras concavum* Zone begins the Induan Stage (Kiparisova and Popov, 1956, 1964; Rostovtsev and Dagys, 1984), the lower one in the Triassic System of the ISS. The Triassic age of the *Otoceras* Beds is accepted by all specialists on ammonoids (Smith, 1932; Spath, 1935; Kiparisova and Popov, 1956, 1964; Kummel, 1957; Rostovtsev and Dagys, 1984; Shevyrev, 1986, 1990).

It should be noted that the *Otoceras* fauna characterizes universally the lower parts of transgressive marine successions unconformably overlying various Permian and older horizons and beginning the Mesozoic historical–geological evolutionary stage. This is obvious in Boreal basins (e.g., in Spitsbergen, eastern Greenland, Arctic Canadian Archipelago, and eastern Verkhoyansk region; Tozer, 1967; Dagys *et al.*, 1986; Embry, 1991; and others) and less distinct in the Peri–Gondwanan regions (Shevyrev and Lozovsky, 1998).

In parallel with study of Triassic deposits, biostratigraphic investigations of Upper Permian marine sections were carried out. The Dorashamian Stage was defined as the terminal Permian unit in the Transcaucasia region (*Osnovnye cherty...*, 1984) and synchronous Changsingian Stage in China (Jin *et al.*, 1994); the latter is now included into the ISS. Until recently, most researchers placed the Permian–Triassic boundary at the base of the *Otoceras woodwardi* Zone and, correspondingly, at the top of the Dorashamian (Changsingian) *Paratirolites kittli* Zone, although the direct succession of both zones has never been observed.[3] Tozer (1988a, 1988b) was first to note that *Paratirolites* and *Otoceras* faunas are separated in space. He explained this phenomenon by a worldwide hiatus corresponding to either the terminal Permian (Dorashamian Age) or the initial Triassic (early Griesbachian time). Prior to the *Otoceras* fauna appearance, the diverse and abundant Paleozoic fauna dies off, and, in opinion of Tozer, this was a worldwide event in the history of oceans, because *"if the beds above the unconformity are obviously synchronous they are entirely suitable as a chronostratigraphic datum"* (Tozer, 1988b, p. 255).

In 1981, the special working group on definition of the global stratotype section and point of the Permian–Triassic boundary (PT GSSP) headed by E. Tozer was organized by the International Subcommission of Triassic Stratigraphy. None of the sections studied by that time met requirements for GSSP definition, because they lack continuous successions of marine deposits, and Tozer perfectly realized this. In 1984, first inquiries of the group members showed that the majority of them (16 of 18) tended to place the Permian–Triassic boundary at the base of the *Otoceras* Zone. The Guryul Ravine section in Kashmir (Nakazawa *et al.*, 1975) was considered as a possible stratotype, but its youngest Permian layers yield paleontological remains inadequate for a definite conclusion.

The situation considerably changed when data on conodont were involved into the biostratigraphic practice. Seeking for reasons, which may explain absence of sections showing the succession of the *Paratirolites* and *Otoceras* beds, some researchers assumed that they are partly or completely synchronous. Moreover, Kozur (1989) and Waterhouse (1976) suggested the Permian

[3] The *Rotodiscoceras–Pleuronodoceras* Zone is now regarded as terminal one in the ammonite scale of the Changsingian Stage. This zone overlies the *Paratirolites (Pseudotirolites)* Beds in China and the Transcaucasia region, where it was identified by Zakharov (1988) who correlated uppermost parts of the Dorashamian and Changsingian stages.

age for the *Otoceras* Beds. First data on conodonts from the Permian–Triassic boundary layers brought unexpected results that seemingly confirmed this assumption. Conodont finds were reported from the *Otoceras* Beds of Himalayas. Their forms were erroneously identified with species known from undoubtedly Upper Permian Dzhulfian–Dorashamian (=Wuchiapingian–Changsingian) beds (Bhatt and Arora, 1984; and others), but this was not confirmed later on (Orchard and Krystyn, 1998).

Trying to prove "synchronism" of the *Otoceras woodwardi* and *Paratirolites* zones, Sweet (1979) compared the Permian–Triassic boundary sections from Kashmir and Salt Range using the formal method of graphic correlation proposed by Shaw. The absurdity of such an approach was so evident that E. Tozer showed, using the same method, that both sections reveal a hiatus between the Permian and Triassic strata, which is diachronous, thus the idea of "synchronism" was fallacious (unpublished letter of 1989 to the members of the Triassic and Permian commissions). Nevertheless, the erroneous idea found a wide recognition among geologists, and the majority of "conodont" adherents became strong enough by that time to vote for pleasing decision. As early as in 1994, only two members of the Working Group on the Permian–Triassic boundary retained their initial opinion of 1984 on the prime significance of the *Otoceras* fauna, and 13 members considered the appearance of conodonts *Hindeodus parvus* as a criterion for defining this boundary. Among adherents of the conodont-based boundary, there appeared soon new members, e.g., Yu.D. Zakharov who defended not long ago the ammonite standard (Kozur *et al.*, 1996).

Chinese professor Yin (1985, 1993) who inherited the Tozer's post of the Chairperson of the aforementioned group suggested to establish the PT GSSP in the Meishan section (South China) situated at the southern slope of the eponymous mount in the Changsing County of Zheliang Province. This section is characterized by the continuous succession of marine sediments at the Permian–Triassic boundary. The section is described in detail in several works (Zhao *et al.*, 1981; Sheng *et al.*, 1984; Yin *et al.*, 1996; Shevyrev and Lozovsky, 1998). Its structure is shown in the figure.

Initially, Chinese researchers (Sheng *et al.*, 1984) were of opinion that the Permian–Triassic boundary should be placed at the base of Bed 25 (White Clay), immediately above which Triassic ammonites *Otoceras?* sp. and other forms appear in Bed 26 (Black Clay). These forms are accompanied by the Permian fossils, ammonites *Pseudogastrioceras* and brachiopods *Neowellerella* and *Crurithyris*, included (coexistence of mentioned taxa is known in the Triassic lower strata of other regions). Later on, the Permian–Triassic boundary was placed inside the thin (16 cm) Bed 27 that yielded the conodont species *Hindeodus parvus* representing a member of the "evolutionary morphogenetic lineage" *Hindeodus typicalis—H. latidentatus—H. parvus—Isarcicella turgida—I. isarica* (Kozur, 1990; Yin *et al.*, 1996). The decision formally meets requirements of the global stratotype section and point.

Later on, the above lineage was revised and subsequently regarded either as *Hindeodus latidentatus praeparvus—H. parvus erectus—Isarcicella turgida—I. staeschei—I. isarica* (Kozur, 1996), or as *Hideouts latidentatus—H. parvus erectus—Isarcicella staeschei—I. isarica*, an independent lateral lineage of which corresponds to *H. priscus—H. turgidus* (Wang, 1995, 1996). The final compromise lineage proposed immediately prior to voting for the boundary definition looks as follows: *Hideouts latidentatus—H. latidentatus praeparvus—H. parvus—Isarcicella isarica* (*Proposal...*, 1999). The uncertainty results, first of all, from the different understanding of stratigraphic range of *H. parvus*, the species first established by H. Kozur and Pyatakova in the basal *Claraia* beds where ammonites *Ophiceras* came from.[4] The taxonomic situation around this species is very puzzling (Wang, 1998). Known at present are six its morphotypes and subspecies (*Hideouts parvus erectus, H. parvus parvus, H. parvus anterodentatus, H. parvus meishanensis*, and others). Consequently, when defining the Permian–Triassic boundary, researchers should come to an agreement which form is most appropriate, otherwise, as Wang justly emphasized, we will have six (!) similar levels substantiated by each subspecies or morphotype. In his opinion, the way out of this situation is in a wide discussion of problems related to conodont biostratigraphy and in elaboration of the agreed resolution. We share his viewpoint completely.

The validity of evolutionary conodont lineages, which include *H. parvus* and its subspecies and morphotypes, was seriously criticized (Baud, 1996; Krystyn and Orchard, 1996; Orchard and Krystyn, 1998; Wang, 1998). As it was established in the Spiti section, *H. latidentatus* occurs above *H. parvus* and *Isarcicella staeschei* coexists with *H. parvus*, whereas *I. turgida* appears below the latter in the Changsing locality. Orchard and Krystyn (1998) do not consider *H. latidentatus* sensu stricto as representing an immediate predecessor of *H. parvus*, viewing the former as a representative of a separate lineage that begins with the Late Permian species *H. typicalis*. In their opinion, conodonts from younger layers formally referred to *H. latidentatus* should be identified with *H. praeparvus*. The first occurrence level of *H. parvus* in the middle of Bed 27 of the Meishan section is also doubted. Par example, Mei (1966), the well-known specialist in conodonts, revised the form from Bed 25 (previously determined as *H. latidentatus*) and identified it with *H. parvus*. Later on, Orchard referred this form to *H.* aff. *parvus*

[4] As Wang (1998) noted, two different forms were selected to represent the holotype, and the first of them should be evidently considered as valid. It differs from the form described under the same name from the *Otoceras* Beds of Kashmir (Matsuda, 1981) and naturally leading to confusion.

(Orchard and Krystyn, 1998). In the same bed, the form coexists with first *Neogondolella (Clarkina) meishanensis* and *N. (Cl.)* cf. *tulongensis* (Tian). Combination of shallow-water *Hindeodus* and deep-water *Neogondolella* forms allowed the mentioned authors to define "mixed" zones (*meishanensis–praeparvus* and *meishanensis–parvus*). Some researchers who prefer the first occurrence level of *parvus* forms for candidate of the Permian–Triassic boundary (Kozur, 1998; Wardlaw and Mei, 1998a; 1998b) consider these zones as successively replacing each other with time. Others (Orchard and Tozer, 1997; Krystyn and Orchard, 1996; Orchard and Krystyn, 1998) justly believe that different *Hindeodus* morphotypes (*praeparvus* and *parvus*) can appear, depending on the facies type, at different levels. On the one hand, they may be concurrent to first *N. meishanensis* species (Selong and Meishan? localities). On the other hand (Arctic Canada), *parvus* form appears slightly after the α datum of the *Neogondolella* evolution that marks the appearance of *N. meishanensis* and *N.* ex gr. *teilorae*. In general, major changes in the *Neogondolella* fauna are close to or coincide with (in mixed biofacies) the first appearance of ammonites representing the genus *Otoceras* and associated with conodonts *H. parvus* or *H. praeparvus*, and *each of the latter can serve as a marker of the Permian–Triassic boundary.* Baud (1996) expressed in fact the same standpoint. Recent data on conodont distribution in the continuous succession of Permian–Triassic boundary layers in the Dolomites confirm these relationships. *H. parvus* was previously established here in the uppermost part of the Tesero Oolitic Member above the *H. latidentatus* (=modern *Hindeodus praeparvus*) Zone, and corresponding event was regarded as marking the Permian–Triassic boundary (Kozur, 1989). In the neighboring section, this species is discovered at a significantly lower level, almost at the base of the above member (Nikora and Perri, 1999). These diachronous first-occurrence levels show that the indicated species is unsuitable for definition of the Permian–Triassic boundary.

Despite the aforesaid, the Subcommission of Triassic Stratigraphy absolutely ignored the problems of conodont biostratigraphy and carried out voting, in written form, questioning who agrees to place the PT GSSP at the base of Bed 27c of the Meishan section. The results of voting are published (Gaetani, 2000): 22 of 27 full members (81%; the group consists of 31 persons in total), voted pro and two contra, with one abstention. Maurizio Gaetani, the Chairperson of the subcommission was happy with the results of voting, because the PT GSSP is the first one accepted for the Triassic System. The voting results were supported by the ISC and approved by the IUGS in March of 2001. In August of 2001, a majestic monument was unveiled at site of the boundary stratotype in Meishan (Yin and Tong, 2001).

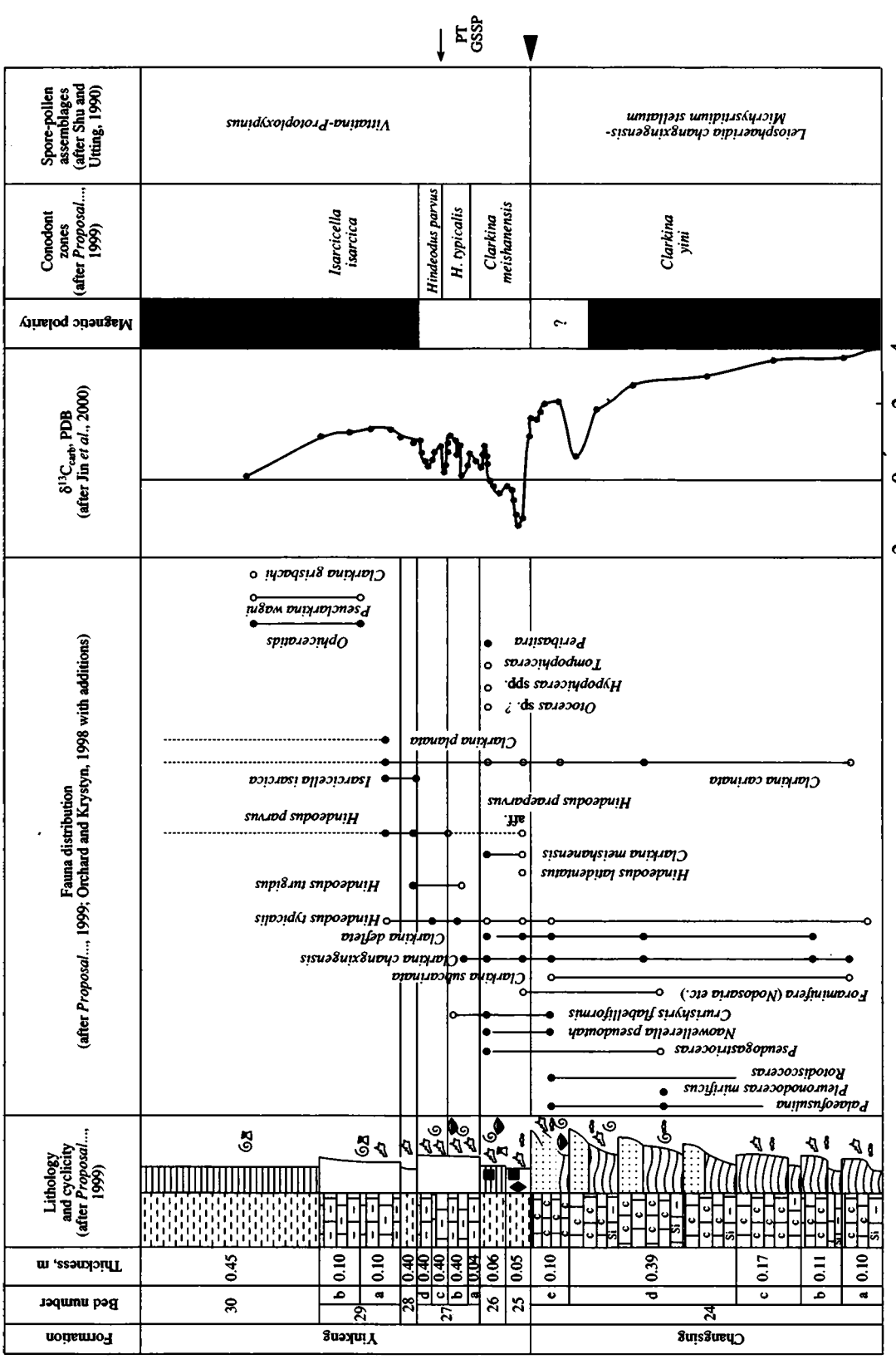
V.R. Lozovsky who took part in voting sent a separate opinion favoring the Meishan section, but indicating that the boundary in question should be placed at the base of Bed 25 in the lowermost Yinkeng Formation conformably overlying the Changsing Limestones. Below, we discuss arguments supporting this standpoint, which were partly mentioned elsewhere (Lozovsky, 1997; Shevyrev and Lozovsky, 1998).

This boundary is easily traceable and appropriate for mapping in South China. It is probably not incidental that Chinese researchers defined previously the Permian–Triassic boundary precisely here (Sheng *et al.*, 1984). The main Permian fauna, including fusulinids (genus *Paleofusulina*), Late Permian ammonoids of the *Rotodiscoceras–Pleuronodoceras* Zone (except for *Pseudogastrioceras* forms having wide stratigraphic range), and many Permian brachiopods, become extinct above Bed 24 although some representatives of the latter group cross the boundary under consideration.

Bed 26 is marked by the appearance of a new assemblage termed by Chinese researchers the *Hypophiceras* fauna (Yin *et al.*, 1996). It includes ammonoids *Otoceras?* sp., *Hypophiceras* cf. *martini* Trömpy, *H. changsingense* Wang, *Metophiceras* sp., *Pseudogastrioceras* sp., and *Tompophiceras* sp. Such a strange composition of the assemblage that includes typical Triassic and one Permian forms is a result, in opinion of Shevyrev (1999, 2000), of their erroneous determinations. From our viewpoint, ammonites determined as *Otoceras?* sp. and associated *Hypophiceras* and close genera rightfully occur at this level. Deposits enclosing this assemblage termed by Chinese scientists (Yin *et al.*, 1996, pp. 49–56) as the *Hypophiceras* fauna can be considered as an important biostratigraphic unit equivalent to the basal part of the *Otoceras woodwardi* Zone of the Spiti section, to the *O. concavum* Zone of boreal basins, and to *Hypophiceras* Zone of eastern Greenland, i.e., to *deposits traditionally regarded as oldest Lower Triassic layers.*[5]

The appearance of boreal *Otoceras* species in central areas of the Tethys is explained by penetration of cold waters from high latitudes, and this event was likely responsible for disappearance of many thermo-

[5] In the Boreal and Tethyan realms, biozonations of the *Otoceras* Beds are based on different *Otoceras* species. In the Boreal realm, this is the universally accepted Tozer's scale consisting of two, the *Otoceras concavum* and *O. boreale* zones. First of these species undoubtedly appears slightly earlier than *O. boreale*, and both species coexist in the middle part of the beds (Henderson and Baud, 1996), whereas only the latter occurs higher passing into the overlying *Ophiceras* Zone. Two zones (*O. latilobatum* and *O. woodwardi*) are defined also in the Tethyan realm.
Mutual correlation between the *Otoceras* zones of both realms is sufficiently reliable, substantiated by a similar evolutionary level of ammonites *Otoceras woodwardi* and *O. boreale*, which are close to each other in both the shell morphology and the lobe line structure. Thus, they can be considered as concurrent (Shevyrev and Lozovsky, 1998). Some researchers (Dagys, 1994; Shevyrev, 2000) believe the *Otoceras concavum* Zone is older than the *Otoceras woodwardi* Zone, whereas some others (Yin, 1993; Orchard and Krystyn, 1998; and others) regard them as synchronous. The latter standpoint well agrees with the data on conodonts published in the last work.



Characteristic features of the Meishan section (South China) where the Global Stratotype Section and Point of the Permian–Triassic boundary is selected: (1) clay; (2) clayey or silty, (3) dolomitized, and (4) organogenic–detrital limestones; (5) volcanogenic admixture; (6) β quartz; (7) horizontal, (8) wavy, and (9) graded bedding; (10) fusulinids; (11) conodonts; (12) brachiopods; (13) bivalve forms; (14) ammonoids; (15) normal and (16) reversed magnetic polarity; (17) approved and (18) proposed variants of Permian–Triassic boundary.

philic Permian organisms (Kozur, 1998). The paradoxical (at first sight) conclusion of Chinese authors on the Late Permian age of the above ammonite assemblage is based mainly on stratigraphic position of the host beds below the accepted boundary corresponding to the first appearance level of conodonts *Hindeodus parvus*, which associate with Permian brachiopods and bivalve forms, two of which are of the Permian age, and the third one represents a new subspecies (*Pteria ussurica variabilis* Chen et Lan) very close to Triassic *Claraia*.[6]

The boundary proposed in the Meishan section is marked also by changes in the conodont fauna used to substantiate the deep-water *Neogondolella* scale: the upper Changsingian *Clarkina yini* Zone is replaced here upward by the *Clarkina meishanensis* Zone (Orchard and Krystyn, 1998; Mei *et al.*, 1999). In the shallow-water zonation of *Hideouts* forms, the lower part of the *meishanensis* Zone corresponds to the *Hideouts praeparvus* Zone overlain by the *H. parvus Zone* mentioned above.[7] In addition to transitional layers of the Meishan section (Beds 25–28), the "mixed" *meishanensis–praeparvus* assemblage is characteristic of the *Otoceras concavum–Otoceras boreale* Zone in Arctic Canada, *Otoceras latilobatum* Zone in Selong, and *Otoceras woodwardi* Zone in Himalayas. These data confirm the ammonite-based correlation mentioned above. The positioning of boundary at the base of Bed 25 and equivalents that accounts for appearance of several forms, including shallow- and deep-water ones, is more reliable than that substantiated by species *H. parvus* only, because their range is understood by different researchers in a different manner, as mentioned above. The additional investigation of conodonts from the Meishan section resulted in discovery of two new *Hindeodus* species appearing in Bed 26 (Metcalfe *et al.*, 2001). These finds once more emphasize significance of the proposed boundary.

The spores–pollen assemblage from the uppermost part of the Changsing Formation of the Meishan section is not diverse (Quyang and Utting, 1990). It includes acritarchs of a wide stratigraphic range and miospores known from the Upper Permian deposits of China. Most interesting components of the assemblage are abundant remains of fungi *Tympanicysta stoschiana* Balme, the bloom of which (the so-called fungal spike) is registered worldwide at the Permian–Triassic boundary both in marine and terrestrial facies and represents an excellent global marker (Wood and Mangerud, 1994; Yaroshenko and Lozovsky, 1997; Lozovsky *et al.*, 2001).[8] The next *Vittatina–Protohaploxypinus* assemblage from overlying layers of the Meishan section (beginning from Bed 25) is of a mixed character because of diverse Late Permian taxa associated at this level with typical Early Triassic species. In general, it is very similar to the *Protohaploxypinus* assemblage from the basal part of the *Otoceras* Beds of eastern Greenland (Balme, 1980) and, to a less extent, to the basal Triassic spore–pollen assemblages of western Canada (Jansonius, 1962) and the Canadian Arctic Archipelago (Utting, 1989). It is considered as the early Griesbachian in age (Shu and Utting, 1990).

Both the *Otoceras* Beds of Arctic Canada (Ogg and Steiner, 1989) and layers yielding the mixed assemblage in China (Yin *et al.*, 1996) correspond in general to paleomagnetic zone N1T of normal polarity. The interval of reverse polarity begins approximately from the level of *Ophiceras* Beds. The normal polarity zone encloses a short-term event of the reverse polarity corresponding to the interval of Beds 25–27 in the Meishan section (Zhu and Lio, 1999) and traceable in oldest continental formations of China (Lozovsky *et al.*, 2001).

The Permian–Triassic boundary in the Meishan section coincides with two major events: the outburst of volcanic activity and replacement of the Late Permian regression by transgression. Some researchers relate the first event to the emplacement peak of traps in the Siberian platform. Wang (1996) defines at the same level also an important lithostratigraphic boundary—the replacement of carbonate sedimentation by the terrigenous–carbonate one. The level is also marked by the negative $\delta^{13}C$ anomaly that used to be explained by a sharp decrease in productivity of marine plankton and mass extinction of organisms.

Even if we leave aside the problem of age of the *Hypophiceras* ammonite fauna and diachronism of the first appearance level of *H. parvus*, any unbiased researcher can clearly see that the boundary between Beds 24 and 25 in the Meishan section marks the beginning of a rapid basin deepening and extensive transgression. It precisely corresponds to the main event boundary within a hiatus separating the undoubtedly Permian deposits from the Triassic strata in the stratotype regions. It separates major stages in development of litho- and biosphere and represents, from the historical–geological standpoint, the most suitable level for the boundary between these systems in the continuous section.

It should be particularly emphasized that the boundary in question is easily traceable in continental sections (*Granitsa permi...*, 1998; Lozovsky *et al.*, 2001). Correlation of continental and marine sections can be done using the basal part of the Tesero Oolitic Member in the Dolomites, where conodonts *Hindeodus praeparvus*, oldest Early Triassic spores–pollen assemblages, and megaspores *Otynisporites eotriasscus* Fugl. occur in association. Two latter associations are regis-

[6] The Triassic forms such as *Claraia* sp. and *Eumorphotis* sp. earlier reported from the same layers (Sheng et al., 1984) are not mentioned in subsequent works.

[7] Kozur (1998) terms it the complex *Clarkina meishanensis–Hindeodus latidentatus praeparvus* Zone.

[8] The negative $\delta^{13}C$ and $\delta^{18}O$ anomalies at the same level are explained by the Late Permian regression and subsequent oxidation of organic matter on the dried shelves. This event increased the CO_2 content in atmosphere and decreased the oxygen content in the ocean–atmosphere system, and both factors favored the mass development of fungi on land and in water basins

tered also at the bases of both the Vetluga Group of eastern Europe and the Buntsandstein of western Europe, the historical stratotype of the Triassic System in the last case. It should be noted as well that the indicated basal levels mark an important structural reorganization in both regions (Lozovsky, 1985).

According to guidelines for GSSP selection (Remane *et al.*, 1996), the priority should be given to the best candidate after discussion of several proposed continuous successions. Considering both variants of the Permian–Triassic boundary in the Meishan section (within Bed 27 that is voted by the majority and at the base of Bed 25 that is proposed here), we arrive at the conclusion that the second one is undoubtedly advantageous. However, the Subcommission on Triassic Stratigraphy did not discuss this variant at all. The hurry in voting not only deserves regret, but also gives birth to doubts in true motives of voting. It is quite clear that a single biostratigraphic criterion used to establish the boundary between systems and based on one evolutionary lineage of conodonts that is under doubts of many scientists, can lead and has already led to impossibility of its definition in many regions, especially when indications of traditional faunas such as ammonoids and historical–geological criteria are completely ignored. For instance, the boundary in question cannot be identified in the vast territory of northeastern Russia, where basal layers of the Lower Triassic transgressive cycle yield the *Otoceras* fauna but are lacking the "guide conodont forms."

Many researches showed that the Triassic continental sequences are as widespread as the Permian ones. The analysis of continental sections of the world (*Granitsa permi...*, 1998) revealed that they, like the marine sections, have a hiatus. In most cases, sequences above the hiatus begin with layers, which can be reliably correlated with the marine *Otoceras* Beds. The Early Triassic age of the layers is well established with the help of different correlation methods (ecotonic, palynological, paleomagnetic).

When the Permian–Triassic boundary is displaced at any higher level, as it happened in the Meishan section, it become completely lost, and this consequence was repeatedly predicted (Lozovsky, 1997, 1998; *Granitsa permi...*, 1998). The present-day situation caused by the resolution of Subcommission on Triassic Stratigraphy is as it was a century ago, when the Paleozoic–Mesozoic boundary strata were termed the "Permotrias." Such a situation is probably convenient for leaders of the International Stratigraphic Commission, but we cannot agree with their decision.

Ovations on occasion of the PT GSSP monument opening still continued, when many researchers abroad became aware of unpleasant consequences of the hasty and unreasoned resolution. Par example, the revision of the Changsing section in the north of the Sichuan Province of China, which is well correlative with the Meishan section based on several stratigraphic markers (White Clay, Black Clay, *Hypophiceras* fauna), showed that conodont species *Hindeodus parvus* occurs in the former five meters higher than in the latter. This led researchers to the conclusion that the Meishan section is possibly lacking a significant "pre-*parvus*" part (Metcalfe *et al.*, 2001). In this case, the established PT GSSP do not meat requirements of the International Stratigraphic Code!

In view of the discussed resolution on the Permian–Triassic boundary, it is necessary to revise the scope of the Griesbachian Stage, the traditional subdivision of the boreal Triassic in Arctic Canada, in which *H. parvus* has been identified for the first time. However, it is hardly possible to do because the accepted boundary does not mark any event (mass extinction, fungal peak, and others) or any physical surface, e.g., a level of sharp facies changes or that of magnetic polarity inversion (Baud, 2001).

CONCLUSION

The discussed examples show that principles used to precise the general stratigraphic scale are not only of the abstract–theoretical interest; they directly influence practice of geological mapping and other geological investigations. The problem is probably not crucial for some countries, but not for Russia with its vast territory insufficiently studied as yet in the geological aspect. For Russia where works on preparation of a new state geological map (Scale 1 : 200000) are in progress, this problem is of a primary importance. If boundaries of all cartographic units do not coincide with boundaries between stages, series, and systems of the general stratigraphic scale, one may ask what such scale is needed for and how it can be correlated with real boundaries and geological history of a particular region. All talks about the necessity of a global chronostratigraphic standard are ruined by geological practice that shows impossibility of tracing the "ideal" levels established in the unique section and based on the first appearance of voluntary (by agreement) selected fossil species.

We are not against the establishment of the Global Boundary Stratotype Sections and Points as one can conclude after reading this paper. The GSSR's are needed as peculiar markers separating intervals of the geological history. In practical geology, this is thought to be a function of the so-called Auxiliary Stratotype Points. However, we are categorically against the hasty "hammered in golden spikes" now common in practice of the International Stratigraphic Commission. The perfection of the general stratigraphic scale representing an idealized section of the upper part of the Earth's crust usable for correlation of particular stratigraphic sections is necessity beyond a doubt. We actively protest against attempts to radically reorganize the traditional scale by means of selecting fully artificial, inci-

dental boundaries of a negligible correlative potential, to refuse the regional–stratigraphic principle used to establish range and boundaries of geological systems, series, and stages, and to ignore the priority principle that guards stability of the general scale.

In this connection, V.A. Krassilov rightfully said : "In practice, they try now to combine chronostratigraphic boundaries with geohistorical events, indications of which are used for correlation. A point stratotype established somewhere near the natural boundary is only a monument in honor of the International Stratigraphic Commission activity. If the sense of chronostratigraphic procedure is elaboration of universal standards, which could serve as a common language for stratigraphers, then such an artificial language resembles Esperanto that is unnecessary as one can see from the experience of international communication. Passion for Esperanto ended long ago because even the imperfect knowledge of a natural language is more useful for international communication than any artificial language" (Krassilov, 1997, p. 7).

In situation when international geological organizations persistently follow the concept accepted under press of country representatives occupying key position in these bodies and do not wish to take into consideration other approaches (a reflection of policy, but not of science), then national geological surveys should decide whether it is reasonable to follow all their resolutions.

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Silicified Microfossils from the Vendian Yudoma Group, the Uchur–Maya Region of Siberia: Facies Dependence and Biostratigraphic Potential

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Received April 27, 2000

Abstract—Microbiota from cherts of the Vendian Yudoma Group, the terminal one in succession of Precambrian deposits of the Uchur–Maya region, southeastern Siberia, is described and analyzed. Dominant in the Yudoma microbiota are remains of morphologically simple filamentous and coccoidal microorganisms, mainly of cyanobacteria. Acritarchs of complex morphology, which have been described from Vendian deposits of some other regions (the so-called Pertatataka assemblage), are lacking in the studied cherts. In its general composition, the Yudoma microbiota resembles other assemblages of Proterozoic silicified microfossils representing remains of morphologically simple microorganisms conservative in the evolutionary aspect. Main factors that controlled its composition were the shallow-water facies affinity of host deposits and taphonomic peculiarities of fossilization, though the Upper Vendian level of host deposits defined by chemostratigraphic data was likely of a some significance as well, because the acritarch assemblages of the Pertatataka type are widespread in the Lower Vendian deposits only. Described in the paper are 20 species of 16 microfossils' genera, the new species *Osculosphaera tarynnachia* sp. nov. included.

Key words: microfossils, cyanobacteria, acritarchs, protists, Vendian, Cambrian, facies, sedimentation settings.

INTRODUCTION

The Yudoma Group is a type subdivision of the Yudomian Complex regarded as the Siberian equivalent of the Vendian, the main candidate for the terminal system of the Precambrian (Sokolov, 1995, 1997, and references therein). Accordingly, any new data on paleontological remains from the group are undoubtedly significant, elucidating biostratigraphic characterization of that system. Microfossils are of prime significance in this case, because they were successfully used during a long period for subdivision of the Vendian type succession in the East European platform (Volkova, 1985, and references therein) and for wider biostratigraphic considerations during the last decade (Zang and Walter, 1992; Knoll, 1992; Zang *et al.*, 1998), when their specific Early Vendian assemblage of the so-called Pertatataka type (termed after the eponymous formation in central Australia) was distinguished.

Unfortunately, silicified microfossils from the Yudoma Group proper are inadequately studied so far. Two works only are devoted to this problem, one characterizing the Yudoma microbiota from two exposures (Lo, 1980), and the other one, in which microfossils found in several separate sections are described (Kolosov, 1982). Lo (1980) systematically described the Yudoma microbiota, but her work was published more than 20 years ago, and, in the consequent period, classification, views on organic nature, and stratigraphic significance of silicified microfossils considerably changed like the understanding of the Yudoma Group structure and stratigraphy. Trying to fill in the blanks in microphytological characteristics of microbiota, I studied in 1990 a series of the group sections most promising the recovery of silicified microfossils. The collected 78 rock samples could bear, judging from field criteria, the fossilized remains of microorganisms. The results of their investigation are presented below.

GEOLOGICAL STRUCTURE AND AGE OF THE YUDOMA GROUP

The Uchur–Maya region, a vast area in southeastern Siberia, is structurally divisible in two zones: the Uchur–Maya plate in the west and the Yudoma–Maya belt in the east. The Riphean and Vendian strata are subhorizontal on the plate, being thicker and deformed into large open folds in the belt. The folds are faulted in addition, so that rock masses are displaced westward along the trust-fault planes (Semikhatov and Serebryakov, 1983). Deposits of the Yudoma Group are widespread within both structural zones. They transgressively, with angular unconformity in places, rest on various horizons of the Riphean sediments and overlap the Archean gneisses of the Aldan shield near the western margin of the Uchur–Maya plate. Everywhere, the eroded upper boundary of the Yudoma succession is

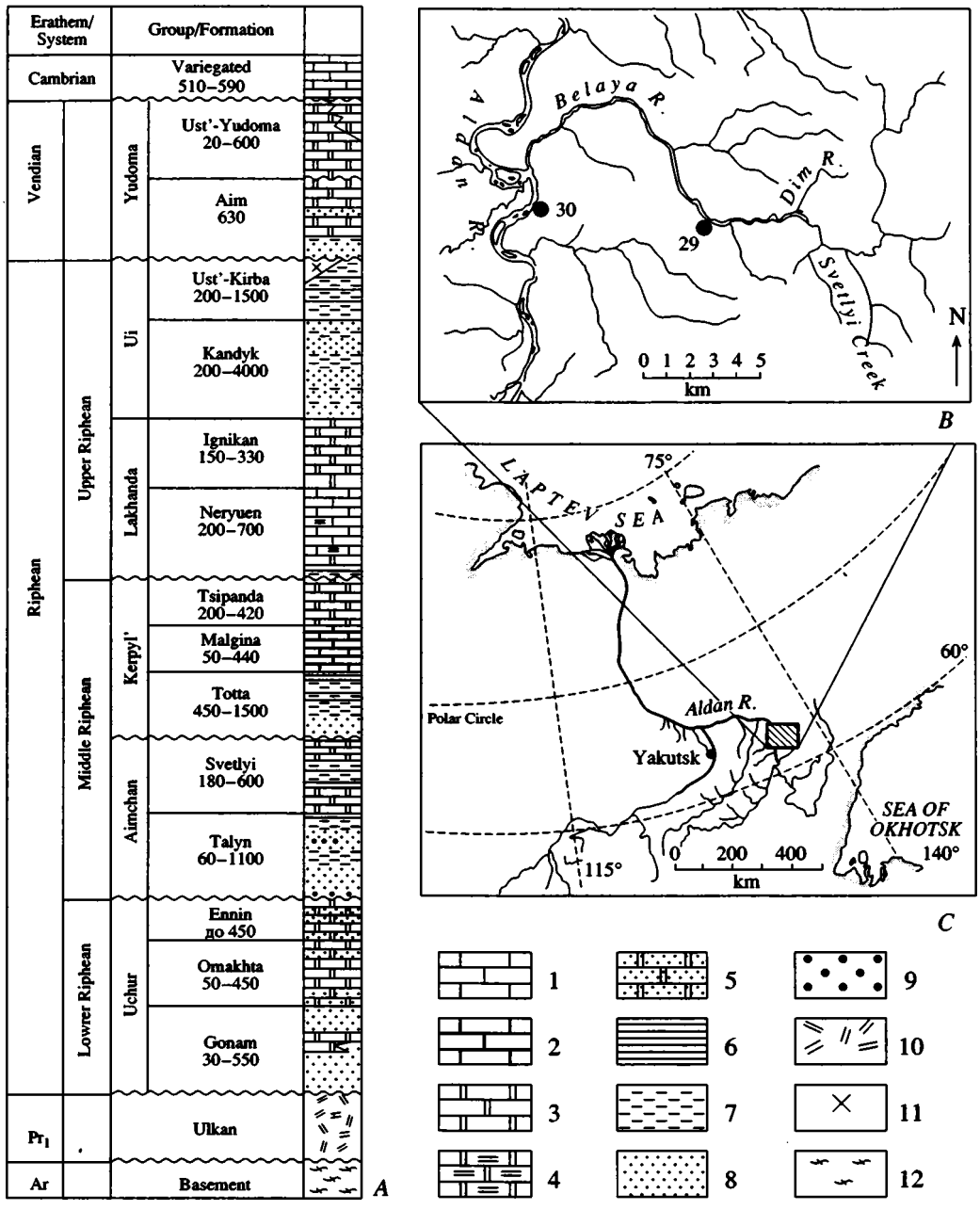


Fig. 1. Stratigraphy (*A*) and geographic localities (*B*, black dots in the inset map) of the Yudoma Group exposures studied in the Belaya and Aldan river basins, the Uchur–Maya region: (1) limestone; (2) variegated algal-laminated limestone; (3) dolomite; (4) alternating limestone, dolomite, and shale; (5) sandy dolomite; (6) shale and claystone; (7) siltstone; (8) sandstone; (9) gravelstone and conglomerate; (10) acidic volcanics and sandstone; (11) ultramafic alkaline intrusions; (12) gneiss and granite gneiss (stratigraphic column after Semikhatov and Serebryakov, 1983; abbreviations in the column: Pr_1, Lower Proterozoic; Ar, Archean).

overlain by variegated deposits of the Lower Cambrian Tommotian Stage.

Lithostratigraphy of the Yudoma Group. The structure, subdivisions, and correlation of the Yudoma group have been exhaustively described in a series of works (Semikhatov *et al.*, 1970; Khomentovskii, 1976; Semikhatov and Serebryakov, 1983; Starnikov *et al.*, 1983; Yakshin, 1984; Sukhorukov, 1984; Yakshin and Pereverzev, 1990; Khomentovskii and Karlova, 1994; and references therein). The group is facies-variable throughout the distribution area, and six types of its rock succession have been originally distinguished (Semikhatov *et al.*, 1970). All successions I have studied in exposures of the Yudoma, Belaya, and Aldan rivers (Fig. 1) are of the Yudoma type and the group hypostratotype. For a rather long period, the group was divided in the lower Aim and upper Ust'-Yudoma formations (Semikhatov *et al.*, 1970; Semikhatov and

Serebryakov, 1983), the lower one used to be thought consisting of sandstone horizons and overlying dark carbonate beds, and the upper one spanning the range of light-colored dolomite beds. Later on, it was established, however, that relationships between both formations are not persistent in different sections of the Uchur–Maya region, and that the Aim formation may decrease in thickness from below, when its basal members turn out to be missing (Semikhatov *et al.*, 1970). Otherwise, this formation can be truncated at the top by the upper dolomitic succession of the Yudoma group (Mezentsev *et al.*, 1978). After the detailed geological survey and prospecting works, the group was divided into the lower Yukanda and upper Sardana formations, the latter unconformably resting on the former with erosion marks in between (Mezentsev *et al.*, 1978; Starnikov *et al.*, 1983; Yakshin, 1984; Sukhorukov, 1984). At the eastern limb of the Gornostakhskii anticline, near the Suordakh Creek mouth, the group was divided into five stratigraphic units: the Nachar, Yalan, Mal, Tokur, and Sytyga formation, the latter four corresponding altogether to the Yukanda Formation. Relationships between all formations distinguished in different sections of the Yudoma group are incompletely clear so far. This problem is under consideration in publications mentioned above, and only those sections are discussed below, in which the microfossils in question have been detected.

The current subdivision of the Yudoma Group is based on the principle of sedimentation cycles that was suggested by Rabotnov and Narozhnykh (1961) who distinguished two cyclothems of the group. Mezentsev and his colleagues (1978) who studied in detail the Yudoma Group exposures in the Allakh-Yun' and Yudoma river basins termed these lithostratigraphic units as the Yukanda (lower) and Sardana formations, the latter resting on the erosion surface of the former. None of the known sections exposes rocks of both formations in their full ranges. Rock successions of the Yukanda Formation are most representative in the east, whereas in the west they are completely cut off by the pre-Sardana disconformity.

A most complete rock succession of the Yukanda Formation is exposed in the Yudoma–Allakh-Yun' interfluve area, where coarse-grained quartz sandstone of its basal bed (less than 1 m thick) grades upward into dolomite that is sandy in the lower part and massive, slightly silicified in the upper part. These rocks underlie the 100-m-thick member of dolomitic limestone beds enclosing interlayers of clayey and oncolitic dolomites, and black cherty lentils. In turn, the member grades upward into a massive dolomite with lentils of light-colored cherts, which is over 100 m thick. The total thickness of the Yukanda Formation below the Sardana sandstones is 260 m (Starnikov *et al.*, 1983). In the western limb of the Gornostakhskii anticline (Exposure 29 in the mouth area of Mutula and Tarynnakh rivers, the Belaya River basin), deposits of the Yukanda Formation that is about 50 m thick are truncated by the pre-Sardana erosion surface. Basal beds of the Yukanda Formation (about 6–8 m thick) are composed in this case of massive quartz–feldspar sandstone, being correlative with the Nachar Formation distinguished near the Suordakh Creek mouth in the eastern limb of the same anticline (Starnikov *et al.*, 1983). They underlie a member of dark gray to black, frequently clayey and microphytoliths-bearing dolomite and limestone beds with intercalations of flat-pebbled conglomeratic breccia, which reveals several levels of dark gray to black cherty lentils and interlayers. Westward, at the Aldan River 12 km upstream of the Belaya River mouth (Kyllakh Ridge, Exposure 30), the Yukanda Formation is completely wedged out, like in the Nuuchalakh site (Exposure 5) of the Yudoma River basin. Only rocks of the Sardana Formation are exposed at both sites.

The most complete rock succession of the Sardana Formation is exposed opposite the Malaya Sakhara River mouth (Maya River basin, western slope of the Gornostakhskii Ridge), where is rests (with erosion marks, though without angular unconformity) on lower beds of the Yukanda Formation. At the base of the Sardana Formation, there is a bed of quartz sandstone about 7.5 m thick, underlying a member of dark gray calcareous dolomite beds of the same thickness, which enclose shale interlayers and are crowned with a bed of quartz sandstone. Above the latter, there is 35-m-thick member of red shales with interlayers of clayey dolomite and red marl. This member underlies the next one composed of gray to dark gray laminated and massive, sometimes calcareous dolomites, which enclose rare interlayers of dark-colored cherts at several levels. The last member is up to 100 m thick, and total thickness of the Sardana Formation is 150 m (Starnikov *et al.*, 1983).

The Sardana Formation is of a similar structure at the Nuuchalakh site (Exposure 5), where its strata truncate the Yukanda beds and directly overlie shales of the Ust'-Kirba Formation (Yakshin, 1984). In the Kyllakh Ridge (Exposure 30 at the Aldan River 12 km upstream of the Belaya River mouth), the Sardana deposits overlie the Tsipanda Formation and reveal again the same succession of beds, although their upper member includes here the dark limestone beds of noticeable thickness, which replace a part of overlying dolomites. In the Belaya River section located in the mouth area of the Mutula and Tarynnakh rivers (Exposure 29, western limb of the Gornostakhskii anticline), the upper part of the Sardana Formation is truncated by the pre-Cambrian erosion (Starnikov *et al.*, 1983).

Age of the Yudoma Group. Age limits are most confidently established for the Sardana and Ust'-Yudoma formations. Small-shelly fossils, which were found first in the former unit by Semikhatov and Serebryakov (1983), are known now from many other levels and imply that the Sardana Formation corresponds almost entirely to the Nemakit-Daldynian (Manykaian) Stage. In Russia, this stage is certified to be the terminal

one in the Vendian System (Sokolov, 1995, 1997), and the Cambrian lower boundary is placed at the base of the first archaeocyathid zone of the overlying Tommotian Stage (*Resheniya...*, 1989). As the standard position of this boundary has been recently fixed at the replacement level of two trace-fossil species (Brasier *et al.*, 1994a; Landing, 1994), which is wittingly pre-Tommotian in the chosen Newfoundland succession, the Nemakit-Daldynian Stage should be included in the Lower Cambrian. The correlation potential of the boundary stratotype in Newfoundland is low however (Rozanov *et al.*, 1997; Khomentovskii *et al.*, 1998), and I retain in my considerations the boundary position at the base of the Tommotian Stage that is traditional for Russian stratigraphers. The U–Pb dating of zircons from volcanic rocks of the Olenek Uplift (northern Siberia) showed that age limits of the Nemakit-Daldynian Horizon correspond to dates of 543 ± 0.24 and 534 ± 0.5 Ma (Bowring *et al.*, 1993; see Rozanov *et al.*, for discussion of stratigraphic position of the dated rocks).

Strict paleontological data, which could elucidate exact time ranges of the Aim and Yukanda formations, are unknown at present, although stromatolites occurring in both subdivisions (Semikhatov *et al.*, 1970) suggest that they are at the Vendian level. Isotopic dates outline a broad time-range for these formations: their upper age limit corresponds to the date characterizing the upper boundary of the Nemakit-Daldynian Stage (see above), and maximum age values, which are measured by the moment in accord with current requirements to validity, range from 942 ± 18 Ma (the Sm–Nd dating method, Pavlov *et al.*, 1992) to 947 ± 7 and 1005 ± 4 Ma (U–Pb dates for basic sills in uppermost Riphean horizons of the Uchur–Maya region, Rainbird *et al.*, 1998).

Available chemostratigraphic data also constrain to some extent the Yudoma Group age. Measurements of carbon isotopic composition in the Ust'-Yudoma Formation showed that $\delta^{13}C$ values vary within its interval from –4 to +6‰ (Magaritz *et al.*, 1986, 1991; Brasier *et al.*, 1994b). Data on carbon isotope composition in rocks of the Khatyspyt, Turkut, and Maastakh formations of the Olenek Uplift, the total range of which corresponds to that of the Yudoma Group (Khomentovskii and Karlova, 1994), depict sharp oscillations of $\delta^{13}C$ values and extend downward the reference curve of secular carbon isotope variations (Knoll *et al.*, 1995; Plechaty *et al.*, 1996a, 1996b). The depicted oscillations typical of the Upper Vendian–Lower Cambrian interval are traceable worldwide (Kaufman and Knoll, 1995; Kaufman *et al.*, 1996; Sailor *et al.*, 1998). The U–Pb dating of zircons from a continuous carbonate succession in southern Namibia was used to establish the time calibration of particular oscillations, and these results constrain the age range of the entire Vendian succession in the Olenek Uplift: the last succession is undoubtedly younger than 550 Ma and corresponds to the Upper Vendian (Sailor *et al.*, 1998).

Thus, the Yudoma Group that bears oldest skeletal fossils of the Nemakit-Daldynian Stage in its upper part is definitely older than 534 ± 0.5 Ma and is younger than 1 Ga, or probably, than 550 Ma.

MICROBIOTA OF THE YUDOMA GROUP

Studied sections and microphytological levels. Analyzing available descriptions of the Yudoma Group structure (Semikhatov *et al.*, 1970; Mezentsev *et al.*, 1978; Semikhatov and Serebryakov, 1983; Starnikov *et al.*, 1983; Yakshin, 1984; Sukhorukov, 1984; Yakshin and Pereverzev, 1990), I understood that sections exposed along the Yudoma, Belaya, and Aldan rivers (in the last case, the Kyllakh Ridge situated 12 km upstream of the Belaya River mouth) are most promising for the discovery of silicified microfossils. Concretions of black, early diagenetic cherts, which usually bear remains of Precambrian silicified microorganisms, are concentrated exactly in these areas. I sampled these concretions from Exposure 29 (site of the Tarynnakh Creek mouth in the Belaya River basin), where groups of samples 4694/35–4694/40 and 4694/41–4694/46 have been collected from intervals the Yukanda and Sardana formations, respectively. The last formation was sampled in addition at Exposure 30 (samples 4694/47–4694/50), in the Kyllakh Ridge situated at the right-hand bank of the Aldan River 12 km upstream of the Belaya River mouth, and at Exposure 5 of the Nuuchalakh site along the Yudoma River (samples 4694/3–4694/7). Microscopic investigation showed that microfossils occur in samples from exposures 29 and 30 (Figs. 1 and 2). Stratigraphic levels of samples are indicated in geological columns (Fig. 2) reproduced from work by Semikhatov *et al.* (1970), in which lithostratigraphy is described in detail. Microfossil assemblages from the Yukanda and Sardana formations do not differ from each other, and they are described below as representing altogether the Yudoma microbiota (Fig. 3).

Composition of microbiota. At lest 20 filamentous and coccoidal microfossil species representing 16 genera have been distinguished in rock samples from the Ust'-Yudoma Formation. Their assemblages from two sampled localities are almost identical. The only distinction is occurrence of endolithic cyanobacteria of the genus *Eohyella* at the right bank of the Aldan River, which are missing from the other site. This seems to be a consequence of microfossils concentration in silicified oolitic interlayers characteristic of the Aldan River site. The Yudoma microbiota includes the following taxa: *Siphonophycus septatum* (Schopf), *S. robustum* (Schopf), *S. typicum* (Hermann), *S. kestron* Schopf, *S. solidum* (Golub.), *Eomicrocoleus crassus* Horodyski et Donaldson, *Eoschizothrix composita* Lee Seong-Loo et Golubic, *Obruchevella* sp., *Archeoellipsoides elongatum* (Golovenoc et Belova), *Eosynechococcus moorei* Hofmann, *Gloeodiniopsis* sp., *Brachypleganon khandanum* Lo, *Sphaerophycus medium* Horodyski et Donaldson, *Eoentophysalis yudomatica* Lo, *Eohyella*

dichotoma Green, Knoll et Swett, *Myxococcoides staphylidion* Lo, *M. chlorelloidea* Knoll, Swett et Mark, *Osculosphaera tarynnachia* sp. nov., *Leiosphaeridia* sp., and two species of acanthomorphic acritarchs *?Meghystrichosphaeridium* sp. and *Micrhystridium* sp. All these taxa represent remains of cyanobacteria and lower unicellular eukaryotes.

Dominant and mat-forming taxa. Dominant forms of the Yudoma microbiota are several *Siphonophycus* species representing empty sheaths of hormogonian cyanobacteria of the so-called LPP (*Lyngbia–Phormidium–Plectonema*) type. According to formal classification of empty *Siphonophycus* sheaths (Butterfield *et al.*, 1994), five species of the genus have been distinguished in the studied microfossil assemblage. These are *S. septatum, S. robustum, S. typicum, S. kestron*, and *S. solidum*. Most abundant among them are empty sheaths *Siphonophycus septatum* (diameter 0.8–2.0 μm), *S. robustum* (diameter 2.0–4.0 μm), and *S. typicum* (diameter 4.0–8.0 μm), which originally formed mats consisting of many dozens of specimens. *S. kestron* and *S. solidum* (diameters 8.0–16.0 and 16.0–24.0 μm, respectively) are less abundant, occurring as individual filaments among colonies of other *Siphonophycus* forms.

It is likely that microbiota under consideration originally included polytrichomatous filaments *Eoschizothrix composita* and *Eomicrocoleus crassus*, i.e., the aggregates of trichomes surrounded by a common sheath, which are comparable with the present-day oscillatorian cyanobacteria of *Schizothrix* and *Microcoleus* genera (Lee Seong-Joo and Golubic, 1998). It is quite difficult to distinguish these cyanobacteria in a fossil state, because the migrated trichomes left, as a rule, only empty sheaths attributable to the form genus *Siphonophycus*. However, one of the colonies preserved in the Yudoma microbiota exhibits a sheath 12.0 μm in diameter that envelopes three trichomes or sheaths having diameters 3.0–3.5 μm (Plate II, figs. 2, 4, 5). Filamentous forms *Eomicrocoleus crassus*, which consist of trichomes 1.2–2.0 μm in diameter set in sheaths having diameters of 15.0–20.0 μm, are comparable with contemporary polytrichomous cyanobacteria of the genus *Microcoleus*, the mat-forming taxon widespread in littoral zones of present-day sea basins (Horodyski and Donaldson, 1980; Hofmann and Jackson, 1991; Kah and Knoll, 1996). However, the fossil clusters of parallel filaments like these may represent remains either of phototrophic prokaryotes, or of nonphototrophic bacteria, and even post-mortem aggregates of monotrichomous cyanobacteria. Accordingly, *Eomicrocoleus* should be regarded as a formal taxon (Lee Seong-Loo and Golubic, 1998). It should be added as well that filaments of *Microcoleus* are of a low taphonomic potential, because sheaths of microorganisms disintegrate so easily that internal trichomes become dispersed (Horodyski *et al.*, 1977; Venetskaya and Gerasimenko, 1988). In a fossil state, the latter would be classified inevitably as *Siphonophycus*

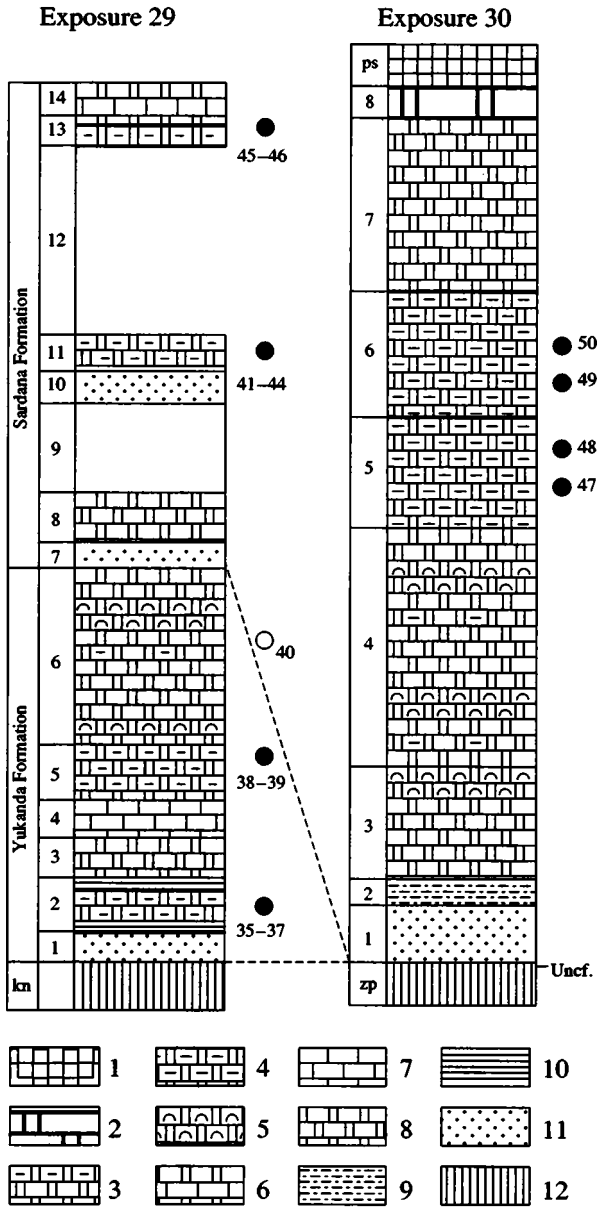


Fig. 2. Lithostratigraphic columns of the Yudoma group, left bank of the Belaya River opposite the Tarynnakh River mouth (Exposure 29), and right bank of the Aldan River 12 km upstream of the Belaya River mouth (Exposure 30): (1) limestone of Variegated Formation; (2) massive to thick-platy dolomite; (3) slightly silicified stromatolitic dolomite; (4) carbonate rocks with cherty lentils and interlayers; (5) dolomite; (6) alternating limestone and dolomite beds; (7) limestone; (8) carbonate rocks with stromatolites; (9) siltstone; (10) shale; (11) sandstone; (12) underlying rocks of the Ust'-Kirba (kn) and Tsipanda (zp) formations (bed numbers after Semikhatov *et al.*, 1970; sample numbers are printed near circles indicating sampling levels). Uncf.—unconformity.

remains (when septae are decomposed), or as certain polytrichomous taxa (when septae are preserved).

Other forms also dominant in the microbiota are cyanobacteria *Eoentophysalis yudomatica*. These

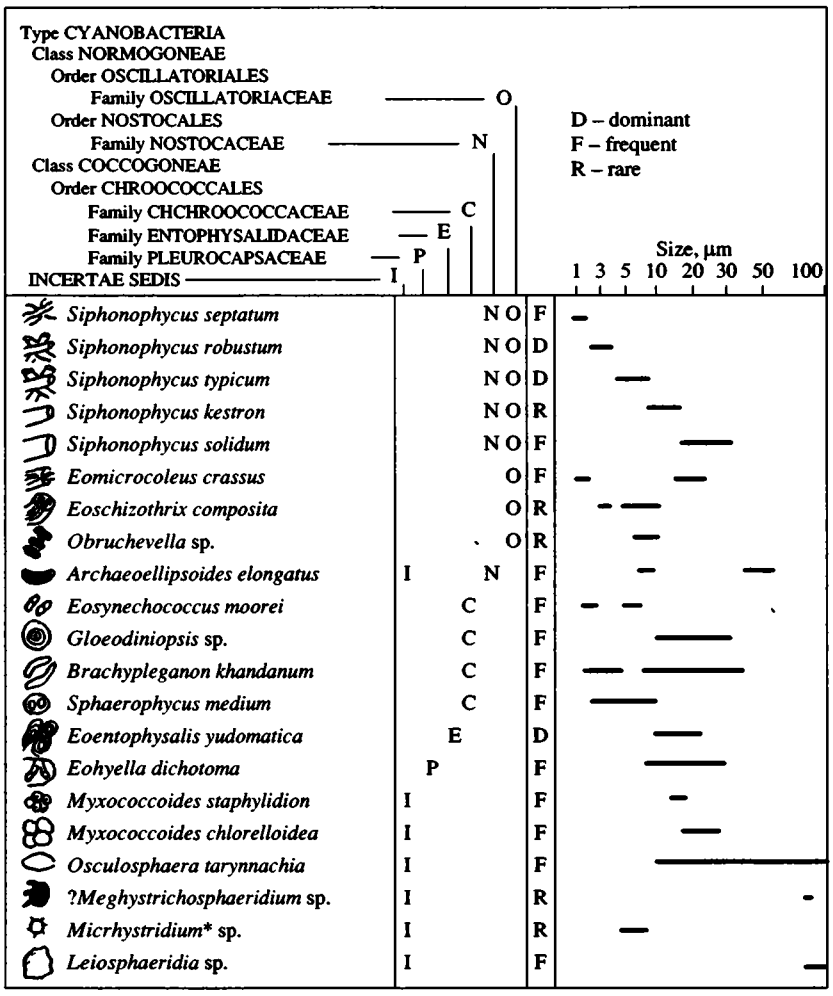


Fig. 3. Synopsis of silicified microfossils distinguished in cherts of the Yudoma Group; size scale is logarithmic; asterisk denotes *Micrhystridium* sp. described by Lo (1980), which have not been detected in studied samples.

multi-layered coccoidal microfossils with external diameter of spheroids ranging from 11.0 to 23.0 μm occur in a form of *Gloeocapsa*-like colonies, which consist of 2, 4, or 8 spheroids surrounded by a common spheroidal envelope, or few dozens of their spheroids may form palmelloid nodular colonies. Another characteristic feature of many Precambrian *Eoentophysalis* forms is their dark coloration in peripheral zones of colonies. Sometimes, spheroids *Eoentophysalis yudomatica* are elongated, showing a unidirectional (polarized) growth (Plate I, figs. 5 and 7) that is their basic morphological feature (Golubic, 1976; Mendelson and Schopf, 1982). In the Yudoma microbiota, colonies of this alga never represent mats consisting of thousands of spheroids. Nevertheless, we certainly deal with this taxon according to the feature mentioned above, but not with chroococcacean cyanobacteria whose secondary aggregations of dead cells are frequently similar in shape (Sergeev and Krylov, 1986; Sergeev, 1992).

Coccoidal symbionts. In colonies of empty sheaths *Siphonophycus robustum* and *S. typicum*, or among their decomposed filaments, there are remains of symbiotic microorganisms, mostly of chroococcacean cyanobacteria, which dwelt in mats of hormogonian blue-green algae. These are spheroidal microfossils *Sphaerophycus medium* (diameter 2.5–10.5 μm), ellipsoidal *Eosynechococcus moorei* and *Brachypleganon khandanum*, smaller and larger axes of which correspond to 2.0–3.0 versus 5.0–7.5 μm and to 2.0–5.0 versus 7.5–37.0 μm, respectively. Remains of these microorganisms are comparable with genera *Chroococcus*, *Synechococcus*, and *Rhabdoderma* of chroococcacean cyanobacteria (Schopf, 1968; Lo, 1980; Butterfield *et al.*, 1994). One more species of the latter, which dwelt in mats of *Siphonophycus* filaments or formed individual colonies, is defined as *Gloeodiniopsis* sp. whose multi-layered envelopes range in diameter from 16.0 to 31.0 μm.

Endolithic microorganisms. The Yudoma microbiota includes remains of endolithic microorganisms *Eohyella dichotoma* rarely occurring in Precambrian deposits and representing peculiar pseudofilamentous

forms 9.0 to 29.0 μm long, which consist of isometric spheroids ranging in diameter from 8.0 to 15.0 μm (size of organic pseudofilamentous structure and recessed burrow) and show a dichotomous or more complicated branching. Like their current (*Hyella* forms from Bahamas) and fossil analogues (e.g., from the Eleonora Bay Group of Greenland; Green *et al.*, 1988), these microorganisms burrowed ooliths and penetrated inside them. Now, they occur in a well-preserved state (Plate I, figs. 1–4). Ooliths, inside which endolithic cyanobacteria *Eohyella rectoclada* dwelt, range in diameter from 50.0 to 100.0–200.0 μm.

Planktonic forms. Remains of eukaryotic and, probably, prokaryotic microorganisms exemplify planktonic forms of the Yudoma microbiota. Among the latter, there are morphologically simple spheroids *Myxococcoides chlorelloidea* and *M. staphylidion,* which have diameters 18.0–27.0 and 14.0–18.0 μm, respectively. These forms may represent remains of chroococcacean cyanobacteria, as it was originally suggested (Schopf, 1968), or of green algae according to morphology of some species (Green *et al.*, 1989). Ellipsoidal microfossils *Archaeoellipsoides elongatus* (7.5–10.0 × 35.0–58.0 μm in size) seem to be spores (akinetes) of planktonic *Anabaena*-like algae. Larger spheroidal microfossils classed with *Leiosphaeridia* sp. (diameter from 75.0 to 175.0 μm) represent remains of unicellular eukaryotic phytoplankton. Dark-colored microfossils *?Meghystrichosphaeridium* sp. (diameter about 75.0 μm) likely exemplify other species of this phytoplankton. They have problematic processes 10.0 to 12.0 μm long, which are of unknown, probable secondary origin.

Peculiar spheroidal microfossils *Osculosphaera tarynnachia* sp. nov., which are fairly abundant in the Yudoma microbiota, usually have elliptic shape in thin sections (15.0–250.0 × 10.0–155.0 μm; Plate II, figs. 4–9). They consist of one or several associated bodies (Plate II, fig. 7) and, according to morphometry, they may represent fossil protists, although their biologic nature is not very clear so far. In opinion of Butterfield *et al.* (1994), the type species *O. hyaline* from the Upper Riphean Baklundtoppen Formation of Svalbard is comparable with some acritarch taxa, in particular, with representatives of the genus *Leiosphaeridia.*

Among microorganisms representing phytoplankton of the Yudoma microbiota, Lo (1980) described small acanthomorphic acritarchs *Micrhystridium* sp., which have diameter 5.0–8.6 μm and processes 1.5 μm long.[1] However, I failed to distinguish the above taxa in the studied material.

[1] Moczydlowska (1991) divided the genus *Micrhystridium* in two genera *Heliosphaeridium* Moczydlowska and *Asteridium* Moczydlowska, species of which have either empty or filled processes, respectively. As it is difficult and even impossible to establish the type of processes for small *Micrhystridium* forms, they are not discriminated in this work.

Other forms. In cherts of the Yudoma Group, I distinguished a poorly preserved specimen of spiral-cylindrical *Obruchevella* sp. Its filament is 7.0 to 11.0 μm in diameter, and diameter of the preserved spiral fragment is about 23.0 μm (Plate II, fig. 8). According to these parameters, the form is close to the species *Obruchevella parva* widespread in Vendian and Cambrian deposits of the Siberian platform and other regions (Sergeev, 1992). The species identification is imprecise however, because that single specimen is inadequately preserved.

In addition to taxa listed above, Lo (1980) described in her work such forms as *Eosphaera?* sp., *Aphetosphaera euthenia*, and branching tubular structures with ellipsoidal inclusions (taxa included as synonyms in description presented below are not mentioned here). I doubt the validity of taxa identified by Lo, but the material available is insufficient for a proper revision.

Environmental position. The initial period of the Yudoma Group accumulation corresponded to the time of a distinct transgression, when sediments were deposited under conditions of an open sea basin, as it is traditionally believed (Semikhatov *et al.*, 1970; Semikhatov and Serebryakov, 1983). Nevertheless, the Yudoma Group, like many other Precambrian successions, accumulated under highly variable sedimentation environments. The concrete environments bearing silicified microfossils are characterized below.

In the section near the Tarynnakh River mouth, microfossils are confined to flat-pebbled conglomeratic breccias, the only silicified components of which usually are pebbles bearing microfossils that set into dolomitic matrix. Breccias of this kind have been interpreted as deposits of either shallow shelf and intertidal areas (Knoll *et al.*, 1991), or deeper settings, where sedimentation developed under influence of storm waves (Knoll *et al.*, 1991; Petrov, 1993). Microfossils occurring in silicified rock fragments imply shallow habitat environments, because their assemblage is lacking taxa characteristic of both the subaerial and proximal shelf settings.

In exposures of the Ust'-Yudoma Formation along the Aldan River, 12 km upstream of the Belaya River mouth, remains of microorganisms have been detected in beds composed of silicified ooliths. Microfossils occur inside the ooliths (endolithic cyanobacteria) or between them, in the last case as individual specimens and colonies. Except for endolithic taxa, the preservation state of microorganisms is poor that is likely a consequence of destruction of nonfossilized microfossils in hydrodynamically active settings.

Oolitic beds of the Ust'-Yudoma Formation (silicified rock varieties included) presumably correspond to deposits of sandy oolitic shoals (Semikhatov *et al.*, 1970), the banks characteristic of coastal sea areas, which isolated shallow settings from open sea areas (Knoll *et al.*, 1991). According to the available data, the above beds obviously represent deposits of an underwater shoal. Cyanobacteria *Eoentophysalis* abundant in

the Yudoma microbiota and most characteristic of the lower intertidal zone could dwell as well in the upper subtidal settings (Petrov *et al.*, 1995). Accordingly, it is very likely that colonies of *Eoentophysalis yudomatica* inhabited the oolitic shoals within the upper subtidal zone of the Yudoma paleobasin, so much more that the same facies yield remains of eukaryotic phytoplankton species *Leiosphaeridia* sp., *Micrhystridium* sp., and others.

Thus, both the taxonomic composition of microfossils and lithologic peculiarities of their host deposits imply that paleoenvironmental location of the Yudoma microbiota corresponded to open marine settings within a shallow inner shelf. Oolitic shoals, which existed at that time, isolated partially or completely some areas of that shelf zone.

BIOSTRATIGRAPHIC SIGNIFICANCE OF SILICIFIED MICROFOSSILS OF THE YUDOMA GROUP

The Yudoma microbiota is of a limited biostratigraphic significance, because it consists of morphologically simple filamentous and coccoidal forms, mostly of cyanobacterial remains. As is acknowledged now, the so-called Pertatataka assemblage is most characteristic one among Vendian microbiotas known from the post-glacial deposits. Its dominant forms are taxonomically diverse and large microfossils with processes and spines, which differ from the Late Riphean and Cambrian taxa. Assemblages of these large acanthomorphic acritarchs are known from cherty-carbonate deposits of the Doushantuo Formation of China (Zhang *et al.*, 1998), from the siliciclastic Pertatataka Formation of central Australia (Zang and Walter, 1992), from silicified dolomites of the Scotia Group (Knoll, 1992), and from the Infrakrol Formation of Lesser Himalayas, India (Tiwari and Knoll, 1994). Similar microfossil assemblages formerly defined as microbiota III (see in Khomentovskii *et al.*, 1987, 1998) have been detected as well in the Moty, Kursovo, Nepsk, Parshino, and other Vendian formations of the Siberian platform (Pyatiletov and Rudavskaya, 1985; Kolosova, 1991; Moczydlowska *et al.*, 1995). Correlation between these deposits and Yudoma Group is ambiguous however, as they have been correlated originally with the Aim Formation (Khomentovskii *et al.*, 1987), but now are thought to be older, pre-Yudoma in age (Khomentovskii *et al.*, 1998). Anyway, only few representatives of the Pertatataka assemblage or microbiota III, such as *?Meghystrichosphaeridium* sp. and *Leiosphaeridia* sp., are distinguishable in the Yudoma microbiota.

The following three factors could be responsible for the peculiar composition of the Yudoma microbiota, in particular, for the absence of representative Pertatataka forms in this assemblage.

(1) The Ust'-Yudoma Formation may be lacking the Pertatataka assemblage because of its relatively younger age. According to current viewpoints, the Doushantuo Formation of China (Zhang *et al.*, 1998), which yield most representative assemblages of large acanthomorphic phytoplanktonic organisms, and Pertatataka Formation of central Australia (Zang and Walter, 1992) are of the Early Vendian age. The Vendian type section of the East European platform (Volkova, 1985) is lacking forms of similar complex morphology and large size, and smaller acanthomorphic acritarchs *Micrhystridium tornatum*, which are comparable in morphometric parameters with *Micrhystridium* sp. described by Lo (1980) from the Yudoma Group, occur in the Kotlin Horizon only. Accordingly, the Ust'-Yudoma Formation may be lacking acritarchs of the Pertatataka type, because its beds bearing silicified microfossils belong to the Nemakit-Daldynian Stage of the Upper Vendian. It is necessary to mention here as well that large acritarchs of the type under consideration have been detected in Vendian deposits of the Irkutsk Amphitheater (Pyatiletov and Rudavskaya, 1985; Kolosova, 1991; and others), and this discovery triggered a long discussion on the Yudoma Formation age.

(2) It is also probable that large acanthomorphic and other acritarchs are missing from the Yudoma Group owing to the shallow-water facies of its beds bearing microfossils. I do not exclude that distribution of the Pertatataka-type taxa was controlled not only by stratigraphic position of their host deposits, but also by the facies affinity of the latter and by paleoclimatic factors. The absence of corresponding taxa in deposits of the East European platform may be not a result of the Lower Vendian hiatus here (in sections of China and Australia, the Pertatataka assemblage occurs immediately below the beds with Ediacarian fauna). This fact could be a possible consequence of sediments deposition in the cold desalinated waters, where development of phytoplanktonic microorganisms characteristic of open marine settings with waters of normal salinity was inhibited (Burzin, 1990, 1994). On the other hand, the shallow-water facies of the Yudoma Group beds bearing microfossils imply that large phytoplanktonic microorganisms were unable to migrate from the seaward shelf areas into sedimentation settings sheltered by oolitic shoals.

(3) One more factor that controlled the taxonomic composition of the Yudoma microbiota could be taphonomic in character. Except for endolithic cyanobacteria preserved in ooliths, all other microfossils occur in oolitic beds in a very poor preservation state. The dead microorganisms obviously had a low chance to remain intact under active hydrodynamic environments thus being intensively deformed. In deposits of the Doushantuo Formation, large acritarchs often represent the growth centers of ooliths (Zhang *et al.*, 1998). One could expect to observe the same situation in rocks of the Yudoma Group, but I failed to find here similar forms, which could be identified for sure.

The above suggestion seems to be reasonable in view of occurrence of the deformed fragments of spiral-cylindrical *Obruchevella* sp. The genus *Obruchevella* is usually widespread in Vendian deposits, but I managed to distinguish in the studied samples only a fragment of corresponding spiral that has been buried between ooliths and may represent a remain of the former large population.

Thus, most of silicified microfossils from the Yudoma Group are characteristic of shallow-water and lagoonal microbiotas of the Riphean and even pre-Riphean age. This fact reflects conservatism of prokaryotic microorganisms, which dwelt in corresponding environmental facies. From older assemblages of microfossils, the Yudoma microbiota differs owing to presence of acanthomorphic acritarchs *Micrhystridium* sp., problematic phytoplanktonic forms *?Meghystrichosphaeridium* sp., and protists remains *Osculosphaera tarynnachia.* Microfossils of this type are known from Upper Riphean and Vendian deposits of different continents. Consequently, the taxonomic composition of the Yudoma microbiota is of a certain stratigraphic significance, indicative of the Late Riphean–Vendian (Neoproterozoic) age of host deposits.

DESCRIPTION OF MICROFOSSILS

Most significant silicified microfossils of the Yudoma Group are described below. They have been studied using the standard investigation procedure of petrographic thin sections in transmitted light. Microscope REM-5 and film Micrat-300 were used to take pictures of specimens whose position was indicated by dots in a paper strap attached to the margin of a given thin section. All the materials studied are stored at the Geological Institute, Russian Academy of Sciences, collection no. 4698.

Kingdom Eubacteria Woese et Fox, 1977

Type Cyanobacteria Stanier *et al.*, 1978

Class Coccogoneae Thuret, 1875

Order Chroococcales Wettstein, 1924

Family Chroococcaceae Nägeli, 1849

Genus *Sphaerophycus* Schopf, 1968

Sphaerophycus medium Horodyski et Donaldson, 1980

Plate II, figs. 12–15

Sphaerophycus medium Horodyski and Donaldson, 1980, pp. 140–141, figs. 5J, K?, L?, M?, N?, 0?, P?, 6C, D; Horodyski and Donaldson, 1983, figs. 5C, D?–H?; Green *et al.*, 1989, fig. 5J; Knoll *et al.*, 1991, p. 553, fig. 19.3; Schopf, 1992, Plate 9, figs. F?–H?; Sergeev *et al.*, 1995, p. 27, figs. 9.6, 9.7.

Sphaerophycus wilsonii Knoll, 1982, pp.783–784, Plate 9, figs. 5–9.

Gloeodiniopsis aff. *gregaria* (partim), Sergeev, 1993, Plate II, figs. 8, 10.

Tetraphycus conjunctum Lo, 1980, pp. 150–152, Plate 3, figs. 1–5.

Description. Monads, dyads, triads, and tetrads of spheroids, which range in diameter from 2.5 to 10.5 μm and are set in a common envelope; the morphotypes usually associate with mat-like colonies of *Siphonophycus* sheaths. Opaque spherical inclusions are frequently present in spheroids.

Remarks. Among microfossils of the Yudoma Group, Lo (1980) described species *Tetraphycus conjunctum* Lo. After the detailed revision, it is found to be identical in its morphometric parameters to *Sphaerophycus medium* thus being regarded here as the younger synonym of the latter.

Distribution. Lower–Middle Riphean: Dismal Lake Group of North America, Kotuikan and Yusmastakh formations of the Anabar Uplift in Siberia; Middle–Upper Riphean: Sukhaya Tunguska Formation of the Turukhansk Uplift in Siberia; Upper Riphean: Draken Conglomerate Formation of Svalbard, Eleonora Bay Group and Limestone-Dolomite "Series" of eastern Greenland; Vendian: Yudoma Group.

Material: several hundreds of specimens.

Genus *Brachypleganon* Lo, 1980

Remarks. Lo (1980) identified the genus *Brachypleganon* as a fossil analogue of present-day *Rhabdoderma*-like chroococcacean cyanobacteria having the elongated, rod-like shape of cells. Cells of this shape discriminate the *Rhabdoderma* species from those of the genus *Synechococcus*, which have ellipsoidal, less elongated cells. The genus *Brachypleganon*, representatives of which are rare in the Precambrian microbiotas, includes only one species.

Brachypleganon khandanum Lo, 1980

Plate II, figs. 13, 14

Brachypleganon khandanum Lo, 1980, p. 156, Plate. II, figs. 9–12; Butterfield *et al.*, 1994, p. 72, figs. 22J, 22K.

Description. These microfossils of elongated elliptical or rod-like shape occur as individual specimens or in aggregates, in which they do not contact with each other. Ellipsoids have the one-layer translucent envelope, the lateral walls of which are of the fine-grained texture, being less than 0.5 μm thick. Longer and shorter axes of ellipsoids range in size from 7.5 to 37.0 and from 2.0 to 5.0 μm, respectively.

Distribution. Upper Riphean: Svanbergfjellet Formation of Svalbard; Vendian: Yudoma Group of Siberia.

Material: about 15 specimens.

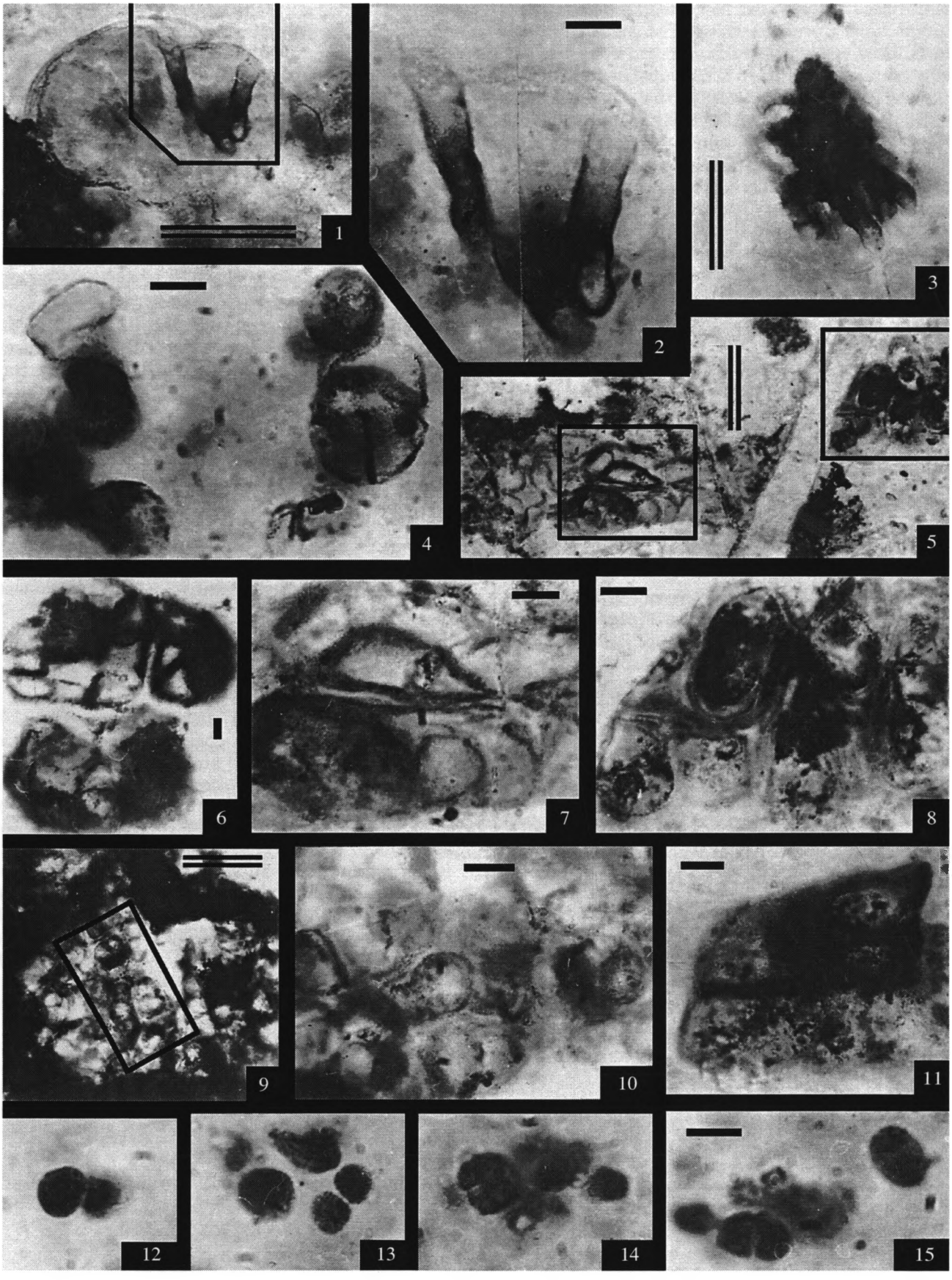
Family Entophysalidaceae Geitler, 1925

Genus *Eoentophysalis* Hofmann emend. Mendelson et Schopf, 1982

Eoentophysalis yudomatica Lo, 1980

Plate I, figs. 5–11

Plate I



Eoentophysalis yudomatica Lo, 1980, pp. 146–150, Plate II, figs. 4–8.

Non *Eoentophysalis yudomatica*: Sergeev and Krylov, 1986, pp. 86–88, Plate IX, figs. 1–4; Yankauskas *et al.*, 1989, p. 90, Plate XXIII, figs. 2, 4, 6; Krylov *et al.*, 1989, Plate I, figs 5a, 5b.

Description. Dyads, tetrads, and octads of multilayer spheroids and colonies consisting of several hundreds of specimens; the colonies are of variable morphology: from "loose, rarefied" clusters of *Gloeocapsa*-like spheroids to palmelloid colonies of irregular or isometric shape. Individual spheroids are frequently ellipsoidal and elongated, presumably demonstrating the accelerated growth of algae in one direction in order to escape burial under sediments. External fine-grained layers of spheroid walls are translucent, about 0.5 μm thick. Similarly translucent, medium- to coarse-grained internal layers of walls are about 1.0 μm thick. Opaque spherical inclusions 0.5 to 3.0 μm in diameter are usually attached to the inner side of the central wall layer. Outer diameters of spheroids (of their external envelope) range from 11.0 to 23.0 μm, and central spherical layer of their envelope is from 10.0 to 22.0 μm across.

Remarks. Coccoidal microfossils described by Lo (1980) as *Eoentophysalis yudomatica* do not occur in a form of corticate colonies consisting of many thousands of cells in contrast to many entophysalidacean algae, but they demonstrate another diagnostic feature of this family—the polarized unidirectional growth (Plate I, figs. 5, 7). Accordingly, these coccoidal microfossils from the Yudoma Group represent remains of Entophysalidaceae but not the occasional post-mortem aggregates of cells of chroococcacean cyanobacteria of the *Gloeodiniopsis lamellosa* type, which occur in the Uralian Upper Riphean Min'yar Formation and initially have been identified as the *Eoentophysalis yudomatica* colonies (Sergeev and Krylov, 1986). Spheroids of *E. yudomatica* are larger in size in distinction from the type species *E. belcherensis*. Their distinction from other large spheroids of the genus *Eoentophysalis*, such as *E. dismallakesensis* (Horodyski and Donaldson, 1980), *E. arcata* (Mendelson and Schopf, 1982), *E. croxfordii* (Muir, 1976), and *E. magna* (McMenamin *et al.*, 1983), which have diameters up to 20 μm and more, are unclear. Further investigations may show that these species are synonyms, or that some of them have no affinity with the genus *Eoentophysalis*.

Distribution. Vendian, Yudoma Group of Siberia.

Material: several hundreds of colonies.

Order Pleurocapsales Geitler, 1925

Family Hyellaceae Borzi, 1914

Genus *Eohyella* Zhang et Golubic, emend. Green, Knoll et Swett, 1988

Eohyella dichotoma Green, Knoll et Swett, 1988

Plate I, figs. 1–4

Eohyella dichotoma Green, Knoll et Swett, 1988, pp. 846–848, figs. 8.5–8.10.

Description. Pseudofilamentous forms penetrating in substratum (mostly in ooliths) and consisting of individual isometric spheroids of irregular ellipsoidal or spherical shape; spheroids have the one-layer envelope and sometimes the dark opaque spherical inclusion. Their walls are in close contact, and some of them are bisected thus giving rise to the false dichotomy of pseudofilaments consisting of spheroids (Plate I, fig. 3). Inside ooliths, structures are frequently disintegrated into separate groups of non-oriented spheroids (Plate I, fig. 4). In other cases, walls between spheroids disappear (a possible result of post-mortem decomposition), and remaining fucoids resemble empty filaments (Plate I, figs. 1, 2). Diameters of spheroids (simultaneously, of pseudofilamentous structures and fucoids) range from 8.0 to 15.0 μm, and their longer axes are from 9.0 to 29.0 μm long. Pseudofilamentous structures and fucoids can be as long as 100–150 μm. In colonies, where pseudofilamentous arrangement of spheroids is indistinct, individual elements range in diameter from 13.5 to 19.0 μm. Envelopes of spheroids 0.5 to 1.0 μm thick are of fine- to medium-grained texture. Inner inclusions range in diameter from 1.0 to 3.0 μm.

Remarks. Four species representing genus *Eohyella* are *E. campbelliae* Zhang et Golubic, *E. rectoclada* Green, Knoll et Swett, *E. endoatracta* Green, Knoll et Swett, and *E. dichotoma* Green, Knoll et Swett. They differ from each other by the character of dichotomy and in morphometric parameters of pseudofilaments and spheroids. Specimens distinguished in the Yudoma Group are most close in morphometric parameters to the species *E. dichotoma*.

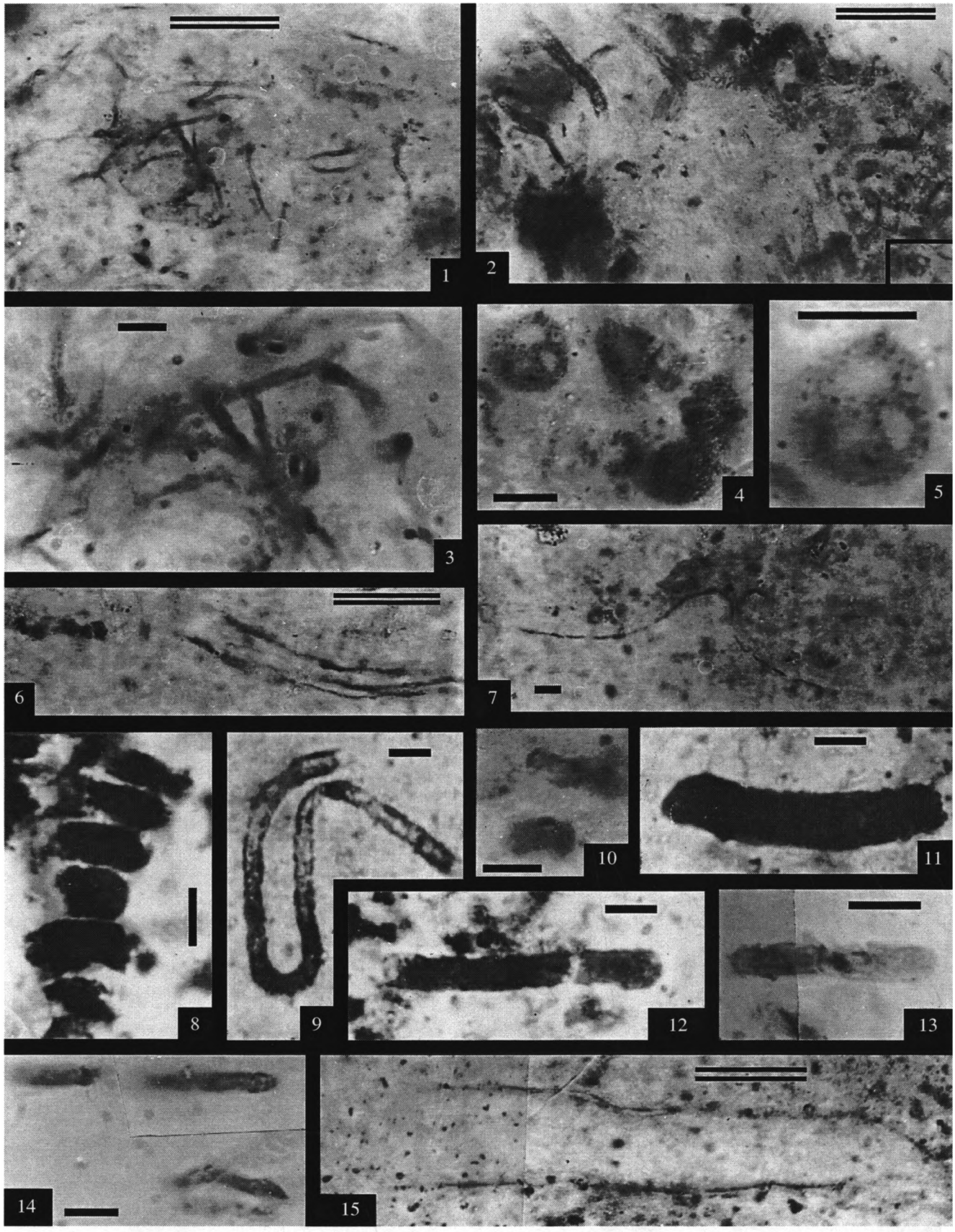
Distribution. Upper Riphean, Eleonora Bay Group of Greenland; Vendian, Yudoma Group of Siberia.

Plate I. Silicified microfossils from the Yudoma group

(1–4) *Eohyella dichotoma* Green, Knoll et Swett: (pentagon in 1, and 2) GIN no. 658, thin section 799, sample 49, t. 2; (3) GIN no. 659, thin section 797, sample 49, t. 1; (4) GIN no. 660, thin section 799, sample 49, t. 1. (5–11) *Eoentophysalis yudomatica* Lo: (left rectangle in 5, and 7) GIN no. 661; (8) (right rectangle in 5, and 8) GIN no. 662, thin section 809, sample 48, t. 2; (6) GIN no. 663, thin section 819, sample 48, t. 4; (clockwise-turned rectangle in 9, and 10) GIN no. 664, thin section 819, sample 48, t. 7; (11) GIN no. 665, thin section 809, sample 48, t. 14. (12–15) *Sphaerophycus medium* Horodyski et Donaldson: (12) GIN no. 666; (13) GIN no. 667, thin section 782, sample 35, t. 1; (14) GIN no. 668; (15) GIN no. 669, thin section 780, sample 35, t. 7.

Samples 48 and 49 are from Exposure 30 (Aldan River), and sample 35 is from Exposure 29 (Tarynnakh River mouth); all from the Sardana Formation. Ordinary and doubled scale bars correspond to 10 and 50 μm, respectively (magnification of figs. 12–14 is identical to that of fig. 15).

Plate II



Material: about 10 colonies, each consisting of several pseudofilamentous structures.

Class Hormogoneae (Geitler) Elenkin, 1934

Order Oscillatoriales Elenkin, 1934

Family Oscillatoriaceae (S.F. Gray) Dumortier ex Kirchner, 1898

Genus *Eoschizothrix* Lee Seong-Joo et Golubic, 1998

Eoschizothrix composita Lee Seong-Joo et Golubic, 1998

Plate II, figs. 2, 4, 5

Eoschizothrix composita Lee Seong-Joo et Golubic, 1998, pp. 181–182, figs. 2, 3, 5, 6, 10

Description. Empty tubular structures of persistent diameter, which are open at both ends and round to oblate in cross section; outer filamentous sheaths enclose one to three of such structures. Hundreds of filaments are clustered into colonies of stratiform or spherical morphology. Diameters of individual tubular structures range from 3.0 to 3.5 μm, and outer sheaths surrounding them are up to 12.0 μm in diameter.

Remarks. In distinction from the genus *Eomicrocoleus* (see below), Lee Seong-Joo and Golubic (1998) regarded *Eoschizotrix* forms as remains of polytrichomatous phototrophic cyanobacteria, fossil populations of which show morphological peculiarities and tendency of behavior characteristic of present-day algae of the genus *Schizothrix*. According to formal criteria, most of *Eoschizotrix composita* remains from the Yudoma Group should be attributed to the genus *Siphonophycus typicum*.

Distribution. Lower–Middle Riphean, Gaoyuzhuang Formation of China; Vendian, Yudoma Group of Siberia.

Material: one colony of several dozens of filaments.

Genus *Eomicrocoleus* Horodyski et Donaldson, 1980

Eomicrocoleus crassus Horodyski et Donaldson, 1980

Plate II, fig. 6

Eomicrocoleus crassus Horodyski et Donaldson, 1980, p. 154, figs. 15A, 15B

Description. Bundles of parallel trichomes surrounded by a common cylindrical sheath; being compressed after death, trichomes are poorly preserved, and their cross-walls are almost invisible. Lateral walls are of the coarse-grained texture. Diameter of trichomes equals to 1.0–2.0 μm, and surrounding sheaths range in diameter from 15.0 to 25.0 μm. Fine- to medium-grained lateral walls of the latter are about 1.0 μm thick.

Remarks. The genus *Eomicrocoleus crassus* has been described from the Middle Riphean Dismal Lakes Group of Arctic Canada (Horodyski and Donaldson, 1980) as a fossil analogue of present-day polytrichomatous oscillatoriacean cyanobacteria of the *Microcoleus, Hydrocoleum*, and *Schizothrix* types. Morphological features of these microfossils are characteristic however not only of blue-green algal taxa mentioned above, but also of some non-phototrophic bacteria and even of post-mortem clusters of originally monotrichomous forms. That is why Lee Seong-Joo and Golubic (1998) distinguished the genus *Eoschizotrix*, species of which undoubtedly represent, according to morphology and distribution patterns in deposits, the fossil analogues of phototrophic polytrichomous cyanobacteria. As a consequence, they suggested to consider the genus *Eomicrocoleus* as a formal taxon, like it was done with respect to the genus *Siphonophycus* (Knoll *et al.*, 1991).

Distribution. Widespread in Precambrian silicified microbiotas.

Material: about twenty of poorly preserved filaments.

Genus *Obruchevella* Reitlinger, 1948

Obruchevella sp.

Plate II, fig. 8

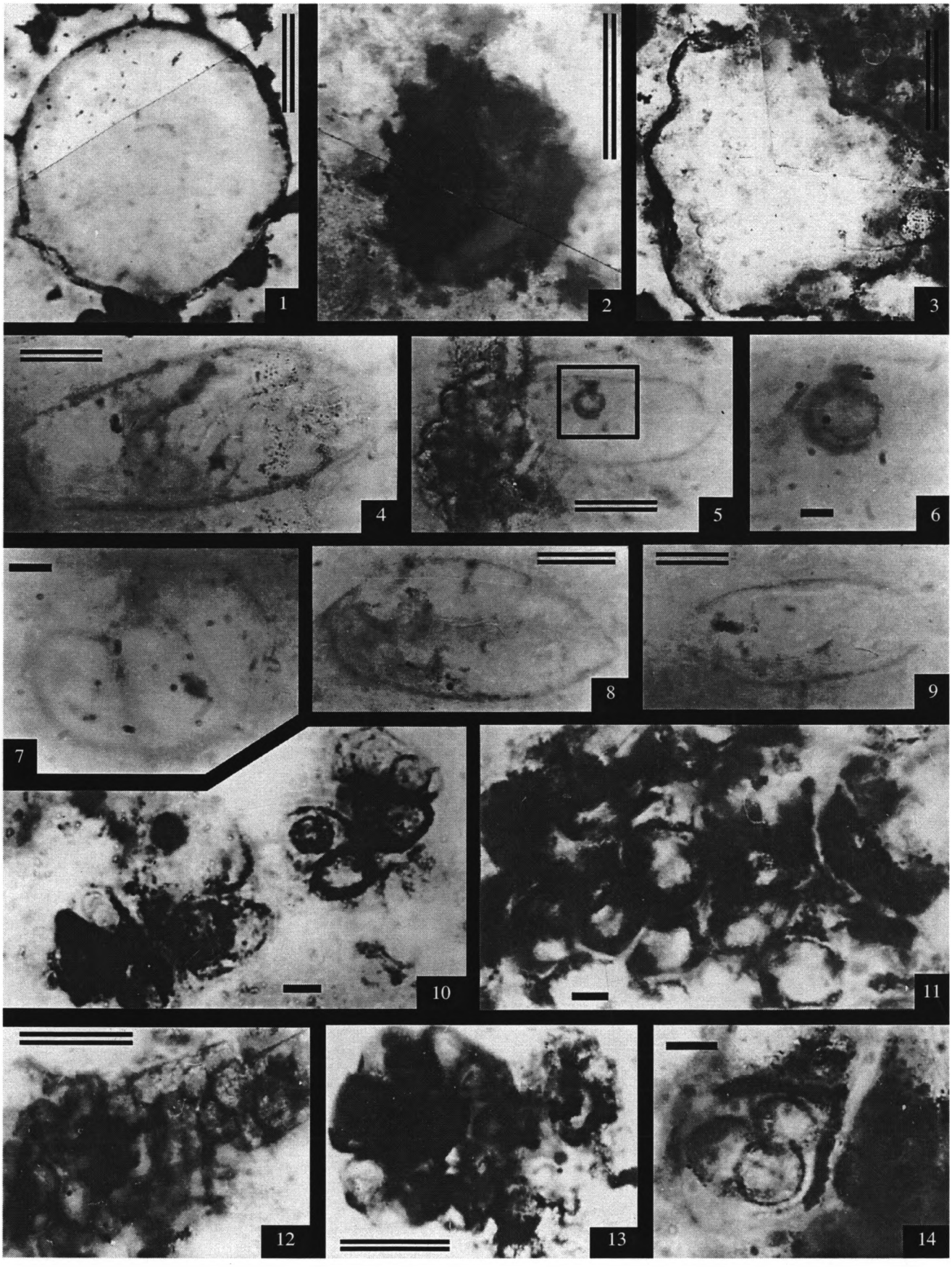
Description. Cylindrical spiral composed of dark opaque filaments; the preserved fragment of spiral is about 23.0 μm in diameter and 56.0 μm long. Diameters of filaments range from 7.0 to 11.0 μm.

Remarks. In its morphometric parameters, the form distinguished in cherts of the Yudoma Group is close to the species *Obruchevella parva*. Its precise taxonomic

Plate II. Silicified microfossils from the Yudoma group

(1, 3) *Siphonophycus robustum* (Schopf), GIN no. 670, thin section 780, sample 35, t. 6. (rectangle in 2, 4, and 5) *Eoschizothrix composita* Lee Seong-Joo et Golubic, GIN no. 671, thin section 773, sample 41, t. 3. (6) *Eomicrocoleus crassus* Horodyski et Donaldson, GIN no. 672, thin section 780, sample 35, t. 5. (7) *Siphonophycus septatum*, GIN no. 673, thin section 780, sample 35, t. 1. (8) *Obruchevella* sp., GIN no. 674, thin section 819, sample 48, t. 9. (9) *Siphonophycus typicum* (Hermann), GIN no. 675, thin section 812, sample 41, t. 2. (10) *Eosynechococcus moorei* Hofmann, GIN no. 676, thin section 820, sample 41, t. 11. (11, 12) *Archaeoellipsoides elongates* (Golovenoc et Belova): (11) GIN no. 677, thin section 818, sample 41, t. 1; (12) GIN no. 678, thin section 812, sample 41, t. 2'. (13, 14) *Brachypleganon khandanum* Lo: (13) GIN no. 679, thin section 826, sample 35, t. 20; (14) GIN no. 680, thin section 826, sample 35, t. 15. (15) *Siphonophycus solidum* (Golub), GIN no. 681, thin section 789, sample 41, t. 1. Samples 35 and 41 are from Exposure 29 (Tarynnakh River mouth), and sample 48 is from Exposure 30 (Aldan River); sample 35 is from Yukanda Formation, and others from the Sardana Formation. Ordinary and doubled scale bars correspond to 10 and 50 μm, respectively.

Plate III



identification is impossible because of the poor preservation state.

Material: one poorly preserved specimen.

INSERTAE SEDIS

Genus *Osculosphaera* Butterfield, 1994

Remarks. To this genus, Butterfield *et al.*, (1994) attributed spheroidal microfossils lacking ornamentation at the surface and having rimmed aperture in their envelopes, diameter of which corresponds to 20–25% of spheroid diameter. These microfossils presumably are remains of eukaryotic phytoplanktonic organisms, though their biological nature is inadequately understood.

Osculosphaera tarynnachia sp. nov.

Plate III, figs. 4–9

The species is termed after the eponymous Tarynnakh River, a tributary of the Belaya River.

Holotype: GIN no. 4698/685; the Uchur–Maya region, Belaya River, Tarynnakh River mouth; Vendian, Yudoma Group.

Description. Individual, coupled, and aggregated spheroids having smooth surface; in aggregates, they are either slightly contacting each other, or set apart. In their morphology, spheroids vary from spherical and vaselake to oblate (ellipsoidal) and fusiform. In the last case, their acute or obtuse ends may be open. Rims around apertures are not very distinct. Envelopes usually consist of one and sometimes of several concentric layers. Some spheroids consist of several (2 to 8) smaller daughter spheroids. The inner envelope layers may be much smaller than the outer ones. Dark spherical inclusions, if present, are attached to internal side of the central layer. Long and short axes of ellipsoidal forms range in size from 15 to 250 and from 10 to 155 μm, respectively. Side walls are amorphous to fine-grained in outer layers (0.5–1.0 μm thick) and medium- to coarse-grained in the inner ones (2.0–3.0 μm thick). Dark spherical internal inclusions have diameters of 2.0–3.0 μm.

Remarks. Ellipsoidal shape of separate spheroids and less distinct rim around apertures differ *Osculosphaera tarynnachia* sp. nov. from *Osculosphaera hyaline* Butterfield. The latter diagnostic features is not clear-cut however, and microfossils under consideration are conventionally attributed to the genus *Osculosphaera.* The new genus has not been distinguished because of absence of necessary distinctive features.

Specimens of *Osculosphaera tarynnachia* likely represent remains of phytoplanktonic microorganisms, as one can judge from the allochthonous character of their distribution in *Siphonophycus* mats. It is likely that separate, ideally shaped spheroids exemplify the lifetime morphology of microorganisms. Being deformed after the death, they gained ellipsoidal outlines, and envelopes of some of them became open at one or both ends. In the last case, further parallel orientation and compression of opposite walls turned them into specimens resembling filaments in thin sections.

Distribution. Vendian, Yudoma Group of Siberia.

Material: about two hundreds of variably preserved specimens.

Genus *Myxococcoides* Schopf, 1968

Myxococcoides chlorelloidea Knoll, Swett et Mark, 1991

Plate III, fig. 11

Myxococcoides chlorelloidea Knoll, Swett et Mark, 1991, p. 558, figs. 13.1–13.4

Description. Individual spheroids with one-layer envelope or their dyads, triads, tetrads (flat and pyramidal), and colonies, in which they are pressed to each other; envelopes are usually translucent, 1.0 to 1.5 μm thick and medium-grained in texture. Diameters of spheroids range from 18.0 to 27.0 μm.

Remarks. At present, the genus *Myxococcoides* includes over 30 species of heterogeneous microfossils, which show overlapping morphometric parameters. Because of simple morphology, it is difficult to discriminate these species from each other. In distinction from other *Myxococcoides* forms, colonies of *M. chlorelloidea* include particular morphotypes resembling certain stages in the life cycle of present-day green algae *Chlorella.*

Plate III. Silicified microfossils from the Yudoma group

(1, 2) *Leiosphaeridia* sp.: (1) GIN no. 682, thin section 766, sample 50, t. 1; (3) GIN no. 683, thin section 773, sample 41, t. 1. (2) *?Meghystrichosphaeridium* sp., GIN no. 684, thin section 790, sample 47, t. 11. (4–9) *Osculosphaera tarynnachia* sp. nov.: (4) holotype, GIN no. 685, thin section 826, sample 35, t. 13; (rectangle in 5 and 6) GIN no. 686, thin section 826, sample 35, t. 13; (7) GIN no. 687, thin section 822, sample 41, t. 1; (8) GIN no. 688, thin section 826, sample 35, t. 18; (9) GIN no. 689, thin section 826, sample 35, t. 18'. (10, 14) *Gloeodiniopsis* sp.: (10) GIN no. 690, thin section 790, sample 50, t. 2; (14) GIN no. 691, thin section 786, sample 50, t. 7. (11) *Myxococcoides chlorelloidea* Knoll, Swett et Mark, GIN no. 692, thin section 790, sample 50, t. 10. (12, 13) *Myxococcoides staphylidion* Lo: (12) GIN no. 693. thin section 834, sample 49, t. 1; (13) GIN no. 681, thin section 834, sample 49, t. 8.

Samples 47, 49 and 50 are from Exposure 30 (Aldan River), samples 35 and 41 from Exposure 29 (Tarynnakh River mouth); sample 35 is from the Yukanda Formation, all others from the Sardana Formation. Ordinary and doubled scale bars correspond to 10 and 50 μm, respectively.

Distribution. Upper Riphean, Draken Conglomerate Formation of Svalbard; Vendian, Yudoma Group of Siberia.

Material: several dozens of specimens.

Myxococcoides staphylidion Lo, 1980

Plate III, figs. 12, 13

Myxococcoides staphylidion Lo, 1980, pp. 152–154, Plate. III, figs. 6–9.

Description. Spheroids with one-layer envelope; they occur in colonies (aggregates) of spherical or irregular shape, which consist of dozens of specimens pressed to each other. Envelopes are usually translucent; external surface and side walls of spheroids about 1.0 μm thick are fine- to medium-grained. Spherical inclusion 1.0 to 2.0 μm in diameter may be attached to internal side of some envelopes. The outer diameters of spheroids range from 14.0 to 18.0 μm.

Distribution. Vendian, Yudoma Group of Siberia.

Material: several specimens.

Genus *Leiosphaeridia* Eisenack, 1958

Leiosphaeridia sp.

Plate III, figs. 1, 2

Description. Isolated spheroids with one-layer translucent envelope; their outlines are round or undulating. External surface and walls of spheroids 2.0 to 3.0 μm thick are fine- to medium-grained. Spheroids range in diameter from 75.0 to 175.0 μm.

Remarks. Some envelopes *Leiosphaeridia* sp. have minute processes at the surface, but critical arguments in favor of primary or secondary origin of these morphological elements are unknown. Specimens of *Leiosphaeridia* sp. are rather frequent in the Yudoma microbiota, and observable distribution patterns suggest their planktonic origin.

Material: few dozens of specimens.

Genus *?Meghystichosphaeridium* Zhang, Yin, Xiao et Knoll, 1998

?Meghystichosphaeridium sp.

Plate III, fig. 2

Description. Single dark opaque envelope with diameter of about 75 μm; it has dark opaque processes of indistinct shape, which are 2.0–4.0 μm in diameter and 7.0–8.0 μm long.

Remarks. This specimen likely represents a remain of phytoplanktonic microorganism. Primary or secondary origin of its processes is unclear however, like in the case of *Leiosphaeridia* sp.

Material: one specimen.

ACKNOWLEDGMENTS

I am grateful to reviewers M.A. Fedonkin and M.S. Yakshin for their constructive comments to the manuscript. The work was supported by the Russian Foundation for Basic Research, project nos. 98-05-64259 and 99-05-64054, and by the American Paleontological Society, grant "PalSirp."

Reviewers M.A. Fedonkin and M.S. Yakshin

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Stratigraphy and Geological Correlation, Vol. 10, No. 6, 2002, pp. 565–574. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 10, No. 6, 2002, pp. 40–49.
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To the Lower Devonian Biostratigraphy in the Russian Far East and Mongolia

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Received June 13, 2000; in final form, May 8, 2001

Abstract—Recent data on the Lower Devonian biostratigraphy in the studied regions are generalized. The described terrigenous and carbonate reference sections yield sharply different faunal assemblages. Carbonate sections of Mongolia, which yield conodont index species of the standard zonation, are correlated with the general stratigraphic scheme of the Devonian. Terrigenous sequences of the Far East and Mongolia are subdivided into regional brachiopod zones of identical stratigraphic succession. The regional zones are calibrated versus the standard Devonian zonations based on their correlation with the carbonate sections.

Key words: Lower Devonian, standard stages, regional biozones, correlation, Mongolia–Okhotsk fauna, endemism, homeomorphs.

Marine shallow-water deposits of the Lower Devonian, which represent terrigenous and carbonate facies, are quite widespread in the Mongolia–Okhotsk region. Predominant and thick terrigenous sequences fill in here the depressions regenerated and newly formed in different periods. They yield abundant and diverse, though endemic fossils attributed to the Mongolia–Okhotsk fauna. This fauna characterizes identical facies in the following region: North Balkhash area of Kazakhstan, Mongolia, Transbaikal and Far East areas of Russia, and northeastern areas of China. Quite different faunal assemblages are characteristic of carbonate deposits in southern Mongolia, where their outcrops are widespread in a latitudinal belt extending from the western to eastern boundaries of the state. Benthic fossil assemblages from these carbonate deposits are comparable with those of the Altai–Sayan region and include remains of conodonts and tentaculites whose correlation potential is very high.

In stratigraphic investigations of the Mongolia–Okhotsk region, the main attention was paid to brachiopods owing to their abundance, diversity, and high evolution ability that enables the detailed subdivision and correlations of sections under study. The progress in investigations was decelerated however by frequent occurrence of homeomorphic brachiopod assemblages. External similarity of brachiopod shells deceived many paleontologists who elaborated stratigraphic and paleobiogeographic schemes based on erroneous identification of brachiopod species. For instance, the Early Devonian basins of the Mongolia–Okhotsk region and North America have been united into a single Pacific paleobiogeographic province, and analogues of North American formations have been erroneously distinguished in the former region (Modzalevskaya *et al.*, 1978; Rzhonsnitskaya *et al.*, 1979).

The comprehensive investigation of brachiopods from Lower Devonian terrigenous deposits of Mongolia showed however that their forms, externally resembling the Appalachian species, differ from the latter in the internal structure and represent other taxa. In addition, the following three important points have been elucidated: (1) brachiopod assemblages from concurrent terrigenous and carbonate deposits differ from each other; (2) among predominant endemic forms, there are species characteristic of the Lochkovian and Emsian stages; (3) terrigenous and carbonate sequences of Mongolia are correlative, respectively, with sections of the Altai Mountains and Salair Ridge. Based on data of intermediate correlation, succession of local stratigraphic units was coordinated with that of standard stages (Alekseeva *et al.* 1981).

We established similarity between brachiopod assemblages from terrigenous deposits of the Far East and Mongolia after revision of available collections. Shishkina who revised the species composition of brachiopods elaborated the detailed stratigraphic scheme of the Devonian in the Far East and correlated it with those of the Altai–Sayan region and Mongolia (Talent *et al.*, 1987; Yolkin *et al.*, 2000). Her scheme was authorized at the 4th Interdepartmental Stratigraphic Conference (*Resheniya...*, 1994). Last year, we begun the detailed reexamination of brachiopods from the Lower Devonian of the Far East and Mongolia (Alekseeva *et al.*, 2001). The data obtained allow us to correlate the terrigenous and carbonate sequences of the Lower Devonian, avoiding the intermediate correlation with concurrent sequences of the Altai–Sayan region. The

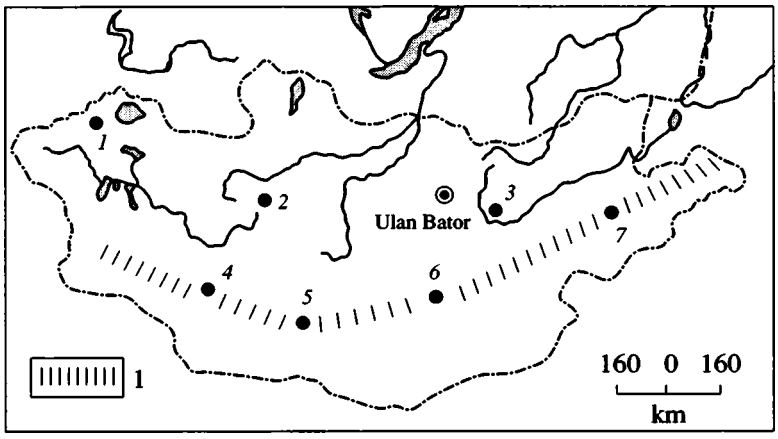

Fig. 1. Geographic localities of the Lower Devonian reference sections studied in Mongolia, those of the carbonate belt (1) included. Sections by names: (*1*) Yamatugol; (*2*) Taryat; (*3*) Kerulen; (*4*) Tsogt-Biger; (*5*) Tsakhirin-Huduk; (*6*) Bortek; (*7*) Barunurt.

investigated Kaila River sections (Amur River middle courses) reveals the continuous succession of Silurian to Middle Devonian beds, which yield faunal assemblages comparable with those characterizing both the terrigenous and carbonate sequences of Mongolia. The latter represent reference sections for the study region, because they yield index species of the standard conodont zones (Alekseeva *et al.*, 1996; Aristov and Nyamsuren, 1998; Vorozhbitov, 1998), and carbonate sequences can be correlated here with the general stratigraphic scale of the Devonian. The most complete carbonate sections with abundant and diverse fauna are known in the latitudinal carbonate belt of southern Mongolia (Fig. 1), where type sections of the Lower Devonian Bortek (Lochkovian), Biger (Pragian), Chulun (Emsian), and Tsaganhalga (upper Emsian–Eifelian) horizons have been distinguished. First three horizons include terrigenous beds intercalated with carbonate deposits. In terrigenous sequences, stratigraphic ranges of horizons are established based on the mutual control of data of interregional correlation with sections of the Altai–Sayan region and on indicative geological events (Alekseeva, 1990). Carbonate and terrigenous facies of aforementioned horizons are separately characterized below. The presented lists of taxa include brachiopod species and some characteristic forms of other associated faunal groups. The complete data on stratigraphic units characterized below and their faunal assemblages are given in synoptic work "Devonian Biostratigraphy of Mongolia" (Alekseeva, 1993).

In carbonate facies, the Bortek Horizon corresponds to the local *Iridistrophia borteki–Howellella angustiplicata* Zone. Its type section is situated in the northern slope of the Paleozoic Bortek massif. In the lower half, the section is composed of tuffaceous conglomerates, green volcanic rocks, and reddish tuffs. They are conformably overlain by dark gray limestone with intercalations of brachiopod coquina and organogenic interlayers, which yield tabulates and stromatoporoids (230 m). Limestones with abundant fauna of similar composition are widespread in the Barunurt massif (Shara-Obi Beds) and Gòby Altai (Amainsajr Formation). Characteristic brachiopods of the horizon are *Mesodouvillina (Protocymostrophia) gobiensis* Mend., *Iridistrophia borteki* Mend., *Stenorhynchia mendae* Erland, *Howellella laeviplicata* (Kozl.), *H. angustiplicata* (Kozl.), *Protathyris praecursor* Kozl., *P. bortegensis* Grunt, *Trematospira alekseevae* Grunt, and *Imacanthyris bortegensis* Grunt. They associate with trilobites *Warburgella rugulosa* (Alth.) and with conodonts *Icriodus woschmidti* Ziegl. and *Ozarkodina reimscheidensis* Ziegl. Species *H. angustiplicata, H. laeviplicata, P. praecursor,* and *W. rugulosa* are widespread in the Lochkovian deposits of Eurasia. First of the conodont taxa is zonal index species in the standard zonation of the Lochkovian Stage, and second one is typical of the Lochkovian deposits in Siberia (Gagiev *et al.*, 1987).

The Biger Horizon of the carbonate belt corresponds to the local *Spirigerina supramarginalis* Zone. It is composed of limestone and thick calcareous conglomerate beds with unsorted pebbles of Silurian rocks. In some sections limestone beds are eroded, and their pebbles occur in basal conglomerate of the Carboniferous. The horizon type section is exposed in southern spurs of the Mongolian Altai along the pass between Biger and Tsogt villages. The horizon is divided here into the Takhiltin and Dutchindaba formation, the former corresponding to a large reef (400 m) composed of gray to light gray algal limestones. Brachiopods found in reefal caverns are *Sieberella praebascuscanica* Kulk., *Leptgonia goldfussiana* (Barr.), *Stenorhynchia nympha* (Barr.), *Wilsoniella prima* Khalf., *Atrypa (A.) gurjevskensis* Aleks., *Spinatrypina margaritoides* Rzon., *S. malobachatica* Aleks., *Carinatina minuta comatoida* Aleks. et Kulk., *Spirigerina supramarginalis* (Khalf.), and *Meristella recta* (Barrios). Their assemblage is similar to that of the Malyi Bachat Horizon of the Salair Ridge.

Limestones of the Takhiltin Formation are conformably overlain by dark gray flaggy limestones of the Dutchindaba Formation (250 m). Organogenic interlayers of the latter are rich in tabulates of a wide stratigraphic range. Tentaculites found in upper layers represent species *Turkestanella acuaria* (Richt.), *T. crenulate* Klish., and *Striatostyliolina parva* Dorodn. et Klish. Two former species are widespread in deposits of the Pragian Stage, and *T. acuaria* represents an index species of the standard tentaculitid zonation (Alberti, 1979; Dorodnova and Klishevich, 1992).

The Chulun Horizon of variable facies is locally composed of terrigenous sediments within the limestone belt. Its type section near the Tsakhirin-Huduk well, the Shine-Dzhinst Village in Goby Altai, is regarded as the reference one for the Emsian Stage, because it reveals the normal succession of the Emsian beds corresponding to the Chulun and overlying Tsaganhalga horizons, which yield the diverse benthic and pelagic faunas. The local *Uncinulus tsakhirinicus* and *Spinatrypa lainae–Cyrtinopsis nalivkini* zones have been distinguished in carbonate facies of the horizon.

Near the base, the *U. tsakhirinicus* Zone is composed of dark gray calcareous shale and sandstone intercalated with subordinate limestone interlayers (5 m). These rocks underlie dark gray medium- to thick-platty limestones with coral interlayers (25 m). Tabulates *Oculipora angulata* Kim and rugoses *Embolophyllum aggregatum aggregatum* (Hill) and *Mycophyllum deficile* Zhelt., which have been identified among corals, are known since the level of the Emsian Stage only. First of these species is typical of the basal Emsian (*regularissimus* Beds) of the Urals, southern Tien Shan, and Altai–Sayan region. The overlying detrital limestones yield the diverse benthic fauna and conodonts. Shale and calcareous sandstone beds with interlayers of organogenic limestone prevail in the upper part of the unit (87 m). Species *Leptagonia zlichovensis, Spinatrypa bachatica*, and *Cyrtina kazi* found here among brachiopods are known only from deposits of the Emsian Stage. Other taxa, such as *Biernatium asiaticus* Aleks., *Lepidoleptaena latiscula* Aleks., *Uncinulus tsakhirinicus* Erlang., and *Cyrtinopsis sursureni* Aleks., are widespread in Mongolia and characteristic of the basal Emsian. Species *Icriodus altaicus* Aks. identified among conodonts is widespread in the Altai Mountains in association with *Polygnathus excavatus gronbergi.* The last taxon was formerly considered as the index subspecies of the second Emsian zone in the standard conodont zonation. This biozone is now termed after *Po. excavatus*, but not after its subspecies mentioned above (Yolkin *et al.*, 1994).

The local *S. galinae–C. nalivkini* Zone includes two, the carbonate and terrigenous members. The lower member (180 m) is composed of dark gray, medium- to thick-platy limestones with biostromes of corals, stromatoporoids, and bryozoans. Conodonts *Po. inversus* Klapp. et Jons, *P. perbonus* (Phill.) *Belodella* sp., and *Pandorinella* sp. occur at the level of 159 m above the member base. The upper beds of detrital crinoidal limestones at the level of 15 m above the conodont-bearing beds yield brachiopods *Levenea orientalis* Aleks., *Leviconchidiella markovi* Aleks., *Latonotoechia* cf. *latona* (Barr.), *Spinartypa galinae* Aleks., *Spinatrypina bachatica* Aleks., *Carinatina arimaspa* (Echw.), and *Cyrtinopsis nalivkini* Rzon. Among them, there are species of the Emsian stratigraphic range and newcomers.

The upper member (152 m) of zone indicated above is composed of greenish tuffaceous and brown terrigenous sandstones and siltstones with lenticular intercalations of sandy crinoidal limestone. The latter yield endemic brachiopod species *Leptaenopyxis zeyensis* Aleks. et Schischk., *Coelospirella dongbeiensis* Su, and *Deltospirifer amurensis* Schischk., which are characteristic of concurrent deposits in the Far East.

The Tsaganhalga Horizon widespread in carbonate facies remains undetected in terrigenous sequences. Deposits of the horizon either transgressively overlap underlying beds different in age, or rest conformably, though with local short-term breaks in sedimentation. In sections of the Goby Altai, their basal portion includes intraformational conglomerates, which are untraceable along the strike. The horizon is divided into the local *Leptodontella zmeinogorskiana* (upper Emsian) and *Leptagonia reticulata* (Eifelian) zones.

Near the Tsakhirin-Huduk well, the *L. zmeinogorskiana* Zone is distinguished above sediments of the Chulun Horizon. At the base of the unit, there is a conglomerate lentil (8 m) with pebbles of underlying rocks. Brownish coarse-grained calcareous sandstones and dark gray limestones (6 m), which rest on the lentil, yield brachiopods *Discomyorthis serotinae* Aleks., *Pseudouncinulus mongolicus* Erlang., and *Coelospirella dongbeiensis* Su. Basal strata of the overlying member of dark gray limestones (40 m) are bearing conodonts *Polygnathus costatus patulus* Klapp., *Icriodus angustus* Sweet, *I. corniger corniger* Welf., *I. corniger leptus* Wedd., and *I. retrodepressus* Bult. The first subspecies is index taxon of the standard Emsian zone. The upper half of the local zone (111 m) is composed of dark gray flaggy and nodular limestones with organogenic intercalations bearing remains of corals, bryozoans, and brachiopods. The identified brachiopod species are *Platyorthis plana* Aleks., *Megastrophia uralensis* (Vern.), *M. (Mesodouvillina) birmanica* (Reed), *Leptodontella zmeinogorskiana* (Peetz.), *L. planuscula* (Khalf.), *Xystostrophia umbraculum* (Schloth.), *Uncinulus shinensis* Erlang., *Eucharitina dobrovi* Erlang., *Elythina salairica* Rzon., and *Undispirifer frequens* Bubl. They associate with tentaculites *Lonchidium gobiense* Dorodn. et Klish, and two new species from genera *Alternatus* and *Dicriconus.* Species predominant among brachiopods are widespread in upper Emsian–Eifelian deposits. The indicated tentaculitid species is similar in morphology to

Tentaculites schlothheimi Koken from the lower Eifelian of Poland (Dorodnova and Klishevich, 1992). The upper boundary of the considered zone is placed at the base of beds bearing conodonts *Polygnathus costatus partitus* Klapp., Ziegl. et Maschk., *Po.* cf. *zieglerianus* Wedd., *Po.* cf. *robustus* Bisch. et Ziegl., *Icriodus* ex gr. *trojani* Johns et Klapp., and *Belodella* sp. First subspecies represents the index taxon of the lower standard zone of the Eifelian Stage.

Let us consider now characteristics of the Lower Devonian terrigenous facies in Mongolia. Exposures of these rocks with abundant faunas are known around the lake Achit Nur in the northwest of the country, where the Bortek Horizon (Lochkovian) corresponds to the Nurskhotgor Formation, the Biger Horizon (Pragian) spans interval of the Tsagansalagol and Turgen formations, and Oshigi and Shanaga beds represent the Chulun Horizon (lower Emsian). The uppermost beds of the Chulun Horizon are widespread in the Khangai Upland. The reference section of Lower Devonian terrigenous facies is exposed along the Yamatugol River, where the continuous succession of beds, which span the interval from the Nurshotgor Formation to the Oshigi beds, i.e., from the Lochkovian to lower Emsian, has been studied.

The Nurskhotgor Formation (100 m) is composed of brownish flaggy inequigranular sandstones of quartz–feldspar composition, which enclose interlayers of nodules, cross-bedded and less abundant calcareous sandstones. Brachiopods from this unit are *Reeftonia borealis* (Hamada), *Schizophoria matutina* Aleks., *Trigonorhynchia tsaganensis* Erlang., *Wilsoniella prisca* Erlang., and *Howellella angustiplicata* (Kozl.). First of these species occurring in the Mongolia–Okhotsk region within the Lower Devonian interval is inadequately studied. *H. angustiplicata* is characteristic of the Lochkovian deposits of Eurasia, and other forms are distinguished at this level for the first time.

The Tsagansalagol Formation (500 m) represents a monotonous sequence of nearly black shales and siltstones with rare interlayers of brownish sandstones and characteristic mass burials of plant remains. Brachiopods *R. borealis* and *Nucleospira musculosa* (Hamada) occurring in this unit have been described from Lower Devonian deposits of the Lesser Hingan.

The Turgen Formation (1050 m) is divided in two subformations. In the lower Turgen Subformation, brownish inequigranular quartz–feldspar sandstones alternate with nearly black shales and siltstones. Dominant rocks of the upper subformation are sandstones. Frequent interlayers of cross-bedded sandstones are characteristic of the entire formation. Shale and siltstone beds are enriched in plant remains. Brachiopods found in sandstones are *Reeftonia taeniolata* (Khalf), *Platyorthis magna* Aleks., *P. carinata* Aleks., *P. alta* Aleks., *Schizophoria kobayashii* (Hamada), *Leptaenopyxis insolita* Aleks. et Schischk., *Gladiostrophia kondoi* (Hamada), *Leptostrophiella kharkraica* (Tchern.), *L. bayrimica* Aleks., *Latonotoechia lata* Erlang., *L. achitensis* Erlang., *Trigonorhynchia ventricosa tarejaensis* Tcherk., *Wilsoniella prima* Khalf., *Spinatrypina margaritoides* Rzon., *Howittia mongoliensis* Aleks. et Mend., *Nucleospira acifera* Aleks. et Mend., *Rhynchospirina convexa* Aleks., *Mongolispira turgensis* Aleks., and *Mongolella altaica* Aleks. Species *R. taeniolata* and *W. prima* are widespread In the Altai–Sayan region within the Pragian–Emsian interval, and subspecies *T. ventricosa tarejaensis* have been described from Emsian deposits of Taimyr. Species *P. magna, S. kobayashii, Lep. insolita, L. kharkraica*, and *S. margaritoides* are characteristic of the Pragian Stage in the Far East and Mongolia.

The basal Oshigi Beds crown the continuous rock sequence exposed along the Yamatugol River. Higher horizons of these and Shanaga beds are exposed in adjacent areas. The Oshigi Beds (260 m) of dark gray shales, sandstones, and greenish siltstones with calcareous sandstone interlayers yield brachiopods *Isorthis (Tyersella) quadrata* Aleks., *Discomyorthis kinsuiensis* (Hamada), *Schizophoria tenuicostata* Aleks., *Leptagonia orientalis* Aleks., *Gladiostrophia kondoi* (Hamada), *Eomaoristrophia kailensis* (Schischk.), *Latonotoechia achitensis* Erlang., *Wilsoniella tchernyshevae* Erlang., *Delthyris grandis* (Kaplun), and *D. khalfini* Aleks. et Mend. Species *S. tenuicostata, L. orientalis, E. kailensis*, and *D. grandis* are characteristic of basal Emsian in the Altai–Sayan region, whereas other forms are widespread in Emsian deposits of the Mongolia–Okhotsk region.

The Shanaga Beds (700 m) of dark gray clayey to sandy limestones alternating with sandstones yield endemic brachiopod taxa *Leptostrophiella depica* Schischk., *Lepidoleptaena latiscula* Aleks., *Ctenochonetes raissae* Afan., *Trigonorhynchia oshigiensis* Erlang., *Coelospirella costata* Aleks., and *Howellella dansani* Aleks. All these species are widespread in basal Emsian strata of the study region.

The Sumyngol Formation represents uppermost strata of the Lower Devonian, which are widespread in the Khangai Upland. Here, at the Taryat site, the formation is composed of dark gray to greenish gravelstones, sandstones, and siltstones with intercalations of nearly black shales (150 m thick in total). The rocks yield brachiopods *Discomyorthis kinsuiensis* (Hamada), *Leptagonia zlichovensis* Havl., *Khangaestrophia raissae* Mend., *Leptostrophiella lata* Aleks., *Rhytistrophia khangaica* Mend., *R. amurensis* Schischk., *R. regularis* Su, *Eomaoristrophia kailensis* Schischk., *E. sumyngolica* Mend., *Trilobostrophia bobilevi* (Schischk.), *Gladiostrophia orientalis* Modz., *G. kondoi* (Hamada), *Areostrophia distorta* (Barr.), *Xenizostrophia* sp., *Suella* sp., *Uncinulus sumynensis* Erlang., *Coelospirella dongbeiensis* Su, *C. chulutensis* Aleks., and *Delthyris grandis* (Kaplun). Predominant in the assemblage composition are endemics of the Mongolia–Okhotsk region, the representatives of genera *Trilobostrophia,*

Xenizostrophia, Suella, and *Coelospirella* included. Species *L. zlichovensis* and *A. distorta* represent taxa of wide geographic ranges. The first of them is known from Emsian deposits, whereas the second one is characteristic of the entire interval of the Lower Devonian.

In the Russian Far East, thick terrigenous and less widespread carbonate–terrigenous sequences with abundant and diverse faunas of the Lower Devonian are known in the Urusha–Ol'doi, Gonzha, and Gag-Sagayan structures of the Mongolia–Okhotsk fold system (Fig. 2). In the Bureya massif, the diverse fossil assemblages have been found in terrigenous sections of the Urmi depression. The same facies of the Nur–Sukhota, Neman–Mel'ga, and some other structural zones are less fossiliferous. In eastern areas of the Far East, the Lower Devonian deposits are of different lithologic type. Rock sequences of predominantly volcanogenic-terrigenous origin yield here rare corals and shells of monotonous composition. Among pelagic taxa, there were found conodonts, stratigraphic significance of which is unclear so far.

The aforementioned areas in upper and middle courses of the Amur River are of a key stratigraphic significance. Shishkina (1990) subdivided here the Bol'shoi Never and Imach horizons of the Lower Devonian into a series of local biostratigraphic units. In the Bol'shoi Never Horizon, these are the *Gladiostrophia pseudofascicula* (Lochkovian), *Leptostrophiella kharkraica* (Pragian), and *Leptostrophiella depica–Delthyris khalfini* (lower Emsian) local zones. In the Imach Horizon, she distinguished the *Khangaestrophia orientalis–Trilobostrophia bobilevi* (middle Emsian) and *Leptodontella zmeinogorskiana* (upper Emsian), and *Fimbrispirifer divaricatus* (Eifelian) local zones. Most complete rock sequences are exposed in middle courses of the Dep River. A normal succession of Silurian to Upper Devonian beds has been studied along the Kaila River, in the southern limb of the Gag syncline. This succession of carbonate–terrigenous rocks represents the reference section of the Lower Devonian in the Far East. Its abundant and diverse fauna is comparable with that from some terrigenous and carbonate sections, and the succession can be also regarded as the reference one for terrigenous facies of the entire Mongolia–Okhotsk region. It has been studied in natural exposures and artificial excavations. The less complete section with abundant fauna of similar composition is exposed along the Obeshchayushchii Creek. Additional data on this section are included in description of the Kaila River section that is presented below.

In the section under consideration, the basal interval (208 m) of the Bol'shoi Never Horizon is composed of quartzite and quartzite-like sandstone beds.

The local *G. pseudofascicula* Zone (110 m) is represented by quartz to quartz–feldspar quartzite-like sandstones with interlayers of clayey limestones and shales. The rocks bear fossil corals, bryozoans, crinoids, and brachiopods. Species identified among the latter are

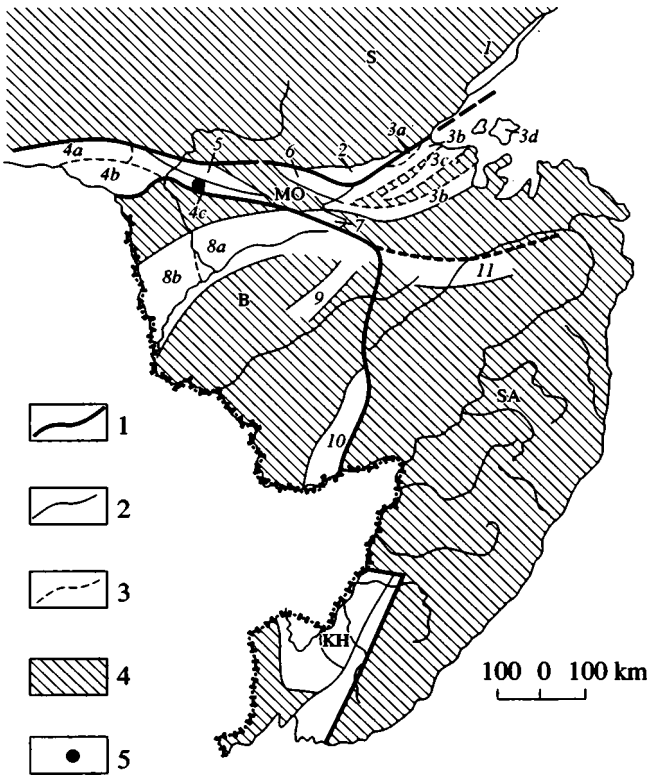

Fig. 2. Structural zones with Devonian deposits in the Russian Far East: (1) principal tectonic boundaries; (2) boundaries of structural zones; (3) other boundaries; (4) areas lacking Devonian deposits; (5) the Kaila River reference section in middle courses of the Dep River.

Principal tectonic structures: (S) Siberian platform; (MO) Mongolia–Okhotsk fold system; (B) Bureya massif; (SA) Sikhote-Alin fold system; (KH) Khankai massif. Structural zones by numbers: (*1*) Ayan; (*2*) Shevla; (*3*) Ud–Shantar and its regions (*3a–3e*); (4) Omutninsk and its Urusha-Ol'doi (*4a*), Gonzha (*4b*), and Gag-Sagayan (*4c*) regions; (*5*) Dolbyr'–Turgala; (*6*) Lan; (*7*) Myn; (*8*) Norsk–Sukhota; (*9*) Niman–Mel'ga; (*10*) Urmi; (*11*) Nilan.

Isorthis (*Protocortezorthis*) *inostranzevi* (Peetz), *Leptaenopyxis lebedevae* (Tchern.), *Tastaria* sp., *Gladiostrophia pseudofascicula* (Rukav.), *Eomaoristrophia carinata* (Borr.), *Eospirifer pseudotogatus* Khalf., *Howellella angustiplicata* (Kozl.), and *Nucleospira musculosa* Hamada. *H. angustiplicata* is known from the Lochkovian deposits of Eurasia, whereas *G. pseudofascicula* and *E. carinata* are characteristic of concurrent sediments in Kazakhstan.

The local *L. kharkraica* Zone (50 m) is composed of quartzitic sandstone and clayey limestone beds, the latter enclosing interlayers of coarse-grained sandstone. Rocks yield corals, bryozoans, bivalves, trilobites, crinoids, and brachiopods. The identified brachiopod taxa are *Isorthis* (*Protocortezorthis*) *inostranzevi* (Peetz), *Platyorthis magna* Aleks., *Schizophoria kobayashii* Hamada, *Leptaenopyxis lebedevae* (Tchern.), *L. insolita* Aleks. et Schischk., *Leptostrophiella kharkraica* (Tchern.), *Iridistrophia eodevonica* Havl., and

Suella chinensis (Hamada). Characteristic of this level in the study region are *P. magna, S. kobayashii, Lep. insolita*, and *L. kharkraica.*

The local *L. depica–D. khalfini* Zone of quartzitic sandstones with interlayers of clayey and organogenic limestones is 45 (Kaila River) to 210 m (Obeshchayushchii Creek) thick. Rare corals and crinoids found in the unit associate with brachiopods *Reeftonia borealis* Hamada, *R. taeniolata* (Khalf.), *Discomyorthis kinsuiensis* (Hamada), *Leptostrophiella depica* Schischk., *Gladiostrophia kondoi* (Hamada), *Eomaoristrophia kailensis* (Schischk.), *Wilsoniella grandis* Hamada, and *Delthyris khalfini* Aleks. et Mend. *L. depica, E. kailensis, W. grandis*, and *D. khalfini* belong to most characteristic taxa.

The local *K. orientàlis–T. bobilevi* Zone, rocks of which are clayey and organogenic limestones, shales, and polymictic, feldspar–quartz, and quartz sandstones with gravel interlayers, is 80 m thick at the Kaila River and 170 m thick at the Obeshchayushchii Creek. In addition to corals, bryozoans, and trilobites, the unit yields divers brachiopods *Discomyorthis kinsuiensis* (Hamada), *Schizophoria tenuicostata* Aleks., *Leptagonia zlichovensis* Havl., *Leptaenopyxis zeyensis* Aleks. et Schischk., *Khangaestrophia orientalis* (Ianov et Modz.), *Rhytistrophia amurensis* Schischk., *R. regularis* Su, *Gladiostrophia kondoi* (Hamada), *Trilobostrophia bobilevi* (Schischk.), *Eomaoristrophia kailensis* (Schischk.), *Xenizostrophia ninniensis* Schischk., *Spinatrypina bachatica* Aleks., *Coelospirella chulutensis* Aleks., *C. dongbeiensis* Su, and *Deltospirifer amurensis* Schischk. In general, this brachiopod assemblage is characteristic of concurrent terrigenous facies throughout the study region. In the carbonate section of the Goby Altai, *Leptagonia zlichovensis* has been found near the unit base, whereas *Lep. zeyensis, C. dongbeiensis*, and *D. amurensis* occur here at a higher level.

The local *L. zmeinogorskiana* Zone is composed of siltstones with limestone and sandstone interlayers (the unit is 80 m thick at the Kaila River and 110 m thick at the Obeshchayushchii Creek). Predominant corals and crinoids associate in this unit with brachiopods *Leptodontella zmeinogorskiana* (Peetz), *L. acuta* Khalf., *Protodouvillina nivkhinorum* Schischk., and *Xystostrophia umbraculum* (Schloth.). All these brachiopod taxa are characteristic of the upper Emsian–Eifelian interval in the Altai–Sayan and study regions. The upper boundary of the zone is placed at the level of considerable changes in composition of brachiopod assemblages.

REGIONAL ZONES

The regional zones are established based on the vertical succession of identical assemblages of endemic brachiopods from deposits under consideration. Two sections with terrigenous rock facies exhibit the continuous succession of beds spanning the interval of all Lower Devonian stages. These are the Kaila River section in the Far East and Yamatugol section in Mongolia. Correlating these sections, we distinguished the *Howellella angustiplicata* (Lochkovian), *Leptostrophiella kharkraica* (Pragian), *Delthyris khalfini* (lower Emsian), and *Trilobostrophia bobilevi* (middle Emsian) regional zones. Their type sections are in the Kaila River succession of beds. The regional *Leptodontella zmeinogorskiana* Zone (upper Emsian) is distinguishable in carbonate–terrigenous sections of the Far East and in carbonate sections of Mongolia. Its stratotype corresponds to section of synonymous local zone in the Goby Altai.

The *H. angustiplicata* Zone spans the range of the local *G. pseudofascicula* Zone and Nurskhotgor Formation, deposits of which yield its index species widespread in the Lochkovian deposits of Eurasia. In the Far East, the Lochkovian age of mentioned units is evident from occurrence of other taxa known from this level in Kazakhstan. According to correlation between sections at the Yamatugol River and in the Altai Mountains, the upper boundary of the Nurskhotgor Formation of Mongolia corresponds to the base of the Yakushino Beds of the Pragian Stage. Brachiopod taxa characteristic of the zone are *Schizophoria matutina, Gladiostrophia pseudofascicula, Eomaoristrophia carinata, Trigonorhynchia tsaganensis, Wilsoniella prisca*, and *Howellella angustiplicata.*

The *L. kharkraica* Zone is identical to synonymous local zone in the Far East and spans the summary range of the Tsagansalagol and Turgen formations in Mongolia. Species in common from the mentioned subdivisions are *Platyorthis magna, Schizophoria kobayashii, Leptaenopyxis insolita*, and *Leptostrophiella kharkraica.* The zone corresponds to the appearance interval of some transit endemic species of brachiopods, which become abundant in overlying deposits. In the Kaila River section, the lower zonal boundary is at the level of changes in brachiopod assemblages, and the upper one marks the appearance level of species widespread in the Emsian deposits. In Mongolia, like in the Altai Mountains, the lower boundary is traceable at the base of the "black shale" member of the Yakushino beds of the Pragian Stage. The upper boundary is established here at the base of beds, which yield brachiopods characteristic of the basal Emsian in the Altai–Sayan region. In addition to taxa in common mentioned above, characteristic of the zone are the following brachiopod species: *Latonotoechia lata, Howittia mongoliensis, Mongolispira turgensis*, and *Mongolella altaica.*

The *D. khalfini* Zone corresponds to the Oshigi and Shanaga beds in Mongolia and to the local *L. depica–D. khalfini* Zone in the Far East. Transit brachiopod species inherited from underlying deposits, coexist at this level with newly appearing taxa known from the Emsian deposits of other regions. These are *Schizophoria tenuicostata, Leptagonia orientalis, Eomaoristrophia kailensis, Iridistrophia eodevonica, Wilsoniella grandis*, and *Delthyris grandis.* Other characteristic

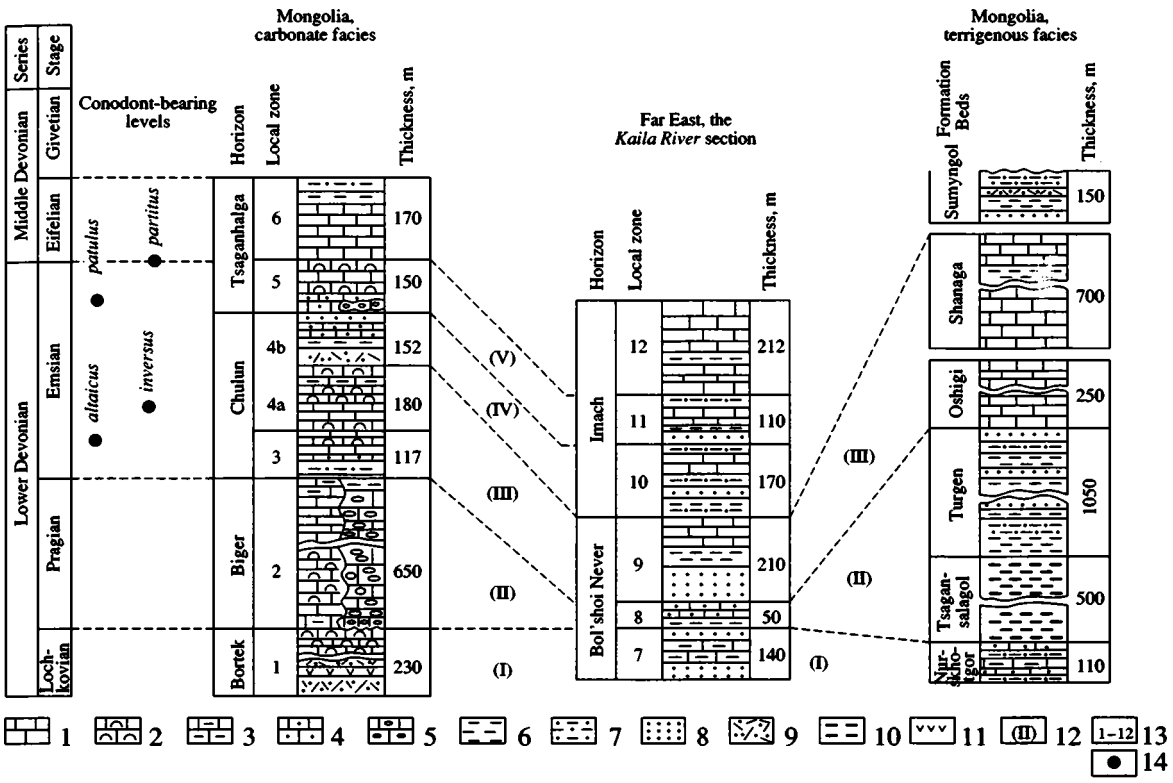

Fig. 3. Correlated Lower Devonian sections of the Far East and Mongolia: (1) limestone; (2) biohermal limestone; (3) clayey limestone; (4) sandy limestone; (5) conglomeratic limestone; (6) shale; (7) siltstone; (8) sandstone; (9) tuffaceous sandstone; (10) "black shales;" (11) lavas; (12) regional zone number; (13) local zone number; (14) conodont-bearing levels.

Local brachiopod zones: (1) *borteki–angustiplicata*; (2) *supramarginalis*; (3) *tsakhirinicus*; (4) *galinae–nalivkini* with lower (4a) and upper (4b) members; (5) *zmeinogorskiana*; (6) *reticulata*; (7) *pseudofascicula–angustiplicata*; (8) *kharkraica*; (9) *depica–khalfini*; (10) *orientalis–bobilevi*; (11) *zmeinogorskiana*; (12) *divaricatus*. Regional zones: (I) *Howellella angustiplicata*; (II) *Leptostrophiella kharkraica*; (III) *Delthyris khalfini*; (IV) *Trilobostrophia bobilevi*; (V) *Leptodontella zmeinogorskiana*.

species of the unit are *Leptostrophiella depica, Ctenochonetes raissae, Trigonorhynchia oshigiensis, Coelospirella costata*, and *Delthyris khalfini*. In the type section, both boundaries of the zone are marked by changes in composition of brachiopod assemblages. In Mongolia, its lower boundary is established at the base of beds that yield species widespread in the basal Emsian of the Altai–Sayan region.

The *T. bobilevi* Zone unites deposits of the local *K. orientalis–T. bobilevi* Zone and Sumyngol Formation. In the Far East and Mongolia, representatives of endemic genera *Khangaestrophia* and *Trilobostrophia* appear for the first time in this interval. Brachiopod assemblages from units under consideration are very similar. Their species in common are *Leptagonia zlichovensis, Khangaestrophia orientalis, K. raissae, Rhytistrophia amurensis, R. regularis, Trilobostrophia bobilevi*, and *Coelospirella chulutensis*. In the type section, both boundaries of the zone are marked by changes in composition of brachiopod assemblages, but in Mongolia they are indistinct.

The *L. zmeinogorskiana* Zone corresponds to synonymous local zones in carbonate–terrigenous sections of the Far East and in carbonate sections of Mongolia. Taxa in common are the index species of the units and *Xystostrophia umbraculum*. Other characteristic species of the zone are *Leptodontella acuta, L. planuscula, Megastrophia uralensis, Elythina šalairica*, and *Undispirifer frequens*. In the type section, the lower zonal boundary is placed at the base of intraformational conglomerates underlying beds with *Polygnathus costatus patulus*, the index species of the upper Emsian zone in the standard conodont zonation. Its upper boundary is at the base of beds with *Po. costatus partitus*, the index species of the lower Eifelian zone in the standard zonation. In the Kaila River section, both boundaries mark changes in composition of brachiopod assemblages.

All regional zones characterizing vast areas in the far East and Mongolia can be considered as the reference biostratigraphic units for the entire Mongolia–Okhotsk region. Their brachiopod assemblages include endemic species, which occur as well in the North Balkhash and Transbaikal regions and in northeastern China. Similarity with the brachiopod fauna from the last region is doubtless. Many species described by Hamada (1971) from the Lesser Hingan are widespread

in the study region. A series of forms pictured in Paleontological Atlases of Northern China (1977, 1980) are also similar to species from the study region. Regretfully, they have been identified with North American taxa based on the external similarity of shell morphology. This erroneous identification with Appalachian species and genera has been already noted by Chinese and American paleontologists (Wang *et al.*, 1984). We did not include these homeomorphic brachiopods in our synopsis because of their brief description and indistinct pictures.

CORRELATION WITH GENERAL STRATIGRAPHIC SCALE

Type sections of four regional zones are established in the Kaila River section correlative with carbonate sections of Mongolia. This offers a possibility to correlate these zones bearing endemic fauna with the general stratigraphic scale of the Devonian (GSSD), because the Mongolian carbonate sections yield index species of the standard conodont zonation, which facilitate such a correlation. Below, we consider first the correlation between the Kaila and carbonate sections, and the distinguishable concurrent subdivisions are correlated afterward with the GSSD.

Terrigenous deposits of the regional *H. angustiplicata* Zone and carbonate sediments of the Bortek Horizon yield a species in common that is acknowledged as the index taxon of the zone. This taxon already defines the Lochkovian age of both subdivisions. In addition, they yield other taxa characteristic of the Lochkovian Stage. In terrigenous rocks of the Kaila section for instance, we found brachiopods *G. pseudofascicula* and *E. carinata* known from the Lochkovian Karazhirik Horizon of Kazakhstan. In carbonate facies of the type and other sections of the Bortek Horizon, the zonal index species coexists with the following representatives of different faunal groups typical of the Lochkovian Stage: brachiopods *Howellella laeviplicata* and *Protathyris praecursor*; trilobites *Warburgella rugulosa*; conodonts *O. reimscheidensis* and *I. woschmidti*. The last one is index species of the lower Lochkovian zone in the standard conodont zonation and enables the direct correlation of the Bortek Horizon with the GSSD.

It is impossible to correlate the regional *L. kharkraica* Zone with carbonate facies of the Biger Horizon based on their fossil assemblages. At the same time, brachiopods from the type section of the horizon are very similar in composition to those from the Malyi Bachat Horizon of the Salair Ridge, where this unit is regarded as an equivalent of the Pragian Stage. The tentaculitid species *T. acuaria* from upper beds of the Biger Horizon is the index taxon of the Pragian Stage in the tentaculitid zonation. The mentioned data allow us to correlate the Biger Horizon with the Pragian Stage. The *L. kharkraica* Zone of the Kaila River section can be correlated with the Biger Horizon according to its position in the section between the Lochkovian and lower Emsian strata. In the Yamatugol section, the Pragian Stage interval is established more precisely than in the Kaila River section. The terrigenous Tsagansalagol and Turgen formations of the former appeared to be correlative with the carbonate–terrigenous Yakushino (Pragian) Beds of the Altai Mountains. The lower boundary of the beds and Tsagansalagol Formation corresponds to the base of the "black shale" member with mass accumulations of plant remains. In the Altai–Sayan region, the upper boundary of the Yakushino Beds is at the base of strata with brachiopods characteristic of the basal Emsian. In the Yamatugol section, the lower and upper boundaries of the Pragian Stage have been controlled by similar events characterizing the transgressive–regressive cycle of the paleobasin evolution. Thus, the regional *L. kharkraica* Zone, like the Biger Horizon, corresponds to the Pragian Stage.

The regional *D. khalfini* and *T. bobilevi* zones of the reference section in Gobi Altai are correlative with local *U. tsakhirinicus* and *S. galinae–C. nalivkini* zones. Deposits of the *D. khalfini* Zone yield brachiopods *Leptagonia zlichovensis* and *Lepidoleptaena latiscula*, which occur as well in the local *U. tsakhirinicus* Zone of the Gobi Altai sections. In addition, lower boundaries of all mentioned subdivisions are established at the base of beds with fossils characteristic of the basal Emsian or first appearing at this level. In terrigenous facies, these are brachiopods *Schizophoria tenuicostata, Leptagonia orientalis, Eomaoristrophia kailensis, Wilsoniella grandis, Delthyris grandis, D. khalfini* and first of the species in common. In the Gobi Altai, the lower boundary of the Emsian Stage corresponds to the base of beds with corals and brachiopods, which are characteristic of or appeared in the basal Emsian strata. These are tabulates *Oculipora angulata*, rugoses *Embolophyllum aggregatum aggregatum, Mycophyllum dificile*, and brachiopods *Spinatrypina bachatica, Cyrtina kazi*, and aforementioned *L. zlichovensis*. The lower limestone member of the local *U. tsakhirinicus* Zone is overlain by beds with conodont species *I. altaicus* that is known in association with *Po. excavatus*, the index species of second Emsian zone in the standard conodont zonation. Consequently, the underlying deposits of *U. tsakhirinicus* and concurrent *D. khalfini* zones may correspond to the lower *Po. kitabicus* Zone in the standard Emsian zonation.

One more level with a comparable brachiopod assemblage corresponds to the regional *T. bobilevi* and local *S. galinae–C. nalivkini* zones. Species in common from this level are *Leptaenopyxis zeyensis, Coelospirella dongbeiensis*, and *Deltospirifer amurensis*. Below this level in the Gobi Altai sections, there are beds with conodonts, among which *Po. inversus* represents the index species of the standard zone. The upper boundary of the *S. galinae–C. nalivkini* Zone is defined at the base of beds with the index subspecies *Po. costatus patulus* of the standard conodont zone. Thus, when cou-

pled, the *U. tsakhirinicus* and *S. galinae–C. nalivkini* local zones, as well as *D. khalfini* and *T. bobilevi* regional zones, are comparable in range with the *Po. kitabicus–Po. serotinus* zonal interval in the standard conodont zonation.

The regional *L. zmeinogorskiana* Zone is correlative with the synonymous local zone in the Gobi Altai section. Species of the genus *Leptodontella* characteristic of both units are widespread and characteristic of the undivided Emsian–Eifelian interval in adjacent areas. In the Gobi Altai, the lower and upper boundaries of the unit correspond to basal levels of beds bearing the conodont subspecies *Po. costatus patulus* and *Po. costatus partitus*, respectively. Accordingly, the regional *L. zmeinogorskiana* Zone spans the range of the standard *Po. costatus patulus* Zone, the uppermost one in the Emsian conodont zonation. The synonymous zone in the Altai–Sayan region is of a wider range and spans the entire upper Emsian Substage (Yolkin *et al.*, 2000).

CONCLUSION

(1) The conodont index species occurring at different levels in Mongolian carbonate sections enable their correlation with the general stratigraphic scale. This situation is used to correlate other stratigraphic subdivisions of local significance with the standard stages of the Lower Devonian.

(2) Terrigenous facies of the Far East and Mongolia, which reveal the identical vertical succession of comparable brachiopod assemblages, are divided into corresponding brachiopod zones of regional significance. The established zonal assemblages are widespread throughout the Mongolia–Okhotsk region.

(3) Type sections of regional zones are correlated with the general stratigraphic scale based on the results of intermediate correlation with concurrent carbonate sections.

ACKNOWLEDGMENTS

We are grateful to A.Yu. Rozanov, the corresponding member of Russian Academy of Sciences, and to M.T. Turbin, the general manager of mining enterprise "Khabarovskgeologiya," and to G.V. Roganov, the chief geologist of the latter, who supervised and helped to organize our investigations. The work was supported by the Russian Foundation for Basic Research, project no. 97-05-64474.

Reviewers E.A. Elkin and A.S. Alekseev

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Candidates for a Global Stratotype of the Induan–Olenekian Boundary (Lower Triassic) in Southern Primor'e

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Received March 10, 2000; in final form, March 28, 2000

Abstract—The Lower Triassic section in the Abrek Bay region, southern Primor'e, is described in detail along with its lithologic, paleontological, and magnetostratigraphic characteristics. This and another section in the Tri Kamnya Cape region, southern Primor'e, are proposed as candidates for the Global Stratotype Section and Point of the Induan–Olenekian boundary. The base of the Olenekian Stage is defined at the appearance level of ammonoid genera *Hedenstroemia* and *Meekoceras*.

Key words: Lower Triassic, Induan and Olenekian stages, biostratigraphy, magnetostratigraphy, southern Primor'e.

Many variants of the Lower Triassic stage subdivision were proposed during the investigation history of Triassic sediments. At the Lausanne Symposium of 1991, the Triassic Subcommission recommended the scheme by Kiparisova and Popov (1964) who suggested to subdivide the Lower Triassic into the Induan and Olenekian stages. At the Triassic Subcommission meeting held in Kyoto during the 29th International Geologic Congress, the subject was discussed again with regard to additional remarks by Baud and Gaetani (1992) and in view of a forthcoming work before the formal ratification of the scheme by the Executive Committee of the International Union of Geologic Sciences. The topical problem indicated by the Triassic Subcommission of the International Stratigraphic Commission is the choosing of a global stratotype of the boundary between the mentioned stages. In this connection, a special working group (IOBWG) chaired by one of the authors (Yu.D. Zakharov) with A. Baud (the Lausanne University) as a vice-chairman was formed in 1997.

The problem of the Induan–Olenekian boundary roots in the fact that none of the sections in the Induan (Salt Range, Pakistan) and Olenekian (Arctic Siberia) stratotype areas (Zakharov, 1996) is appropriate for the boundary stratotype because of the following reasons. (1) The Induan Stage in the Olenek River basin is mainly composed of lagoonal tuffaceous sediments lacking reliable paleontological characteristics. (2) In most sections known in the Indus River basin (Salt Range) and adjacent Himalayan regions, ammonoids are scarce or missing immediately in the boundary beds between the Ceratite Marl and Ceratite Sandstone, where the boundary between the Lower Triassic stages was suggested to be.

Consequently, there is a problem of global correlation of the Lower Triassic stages, which are distinct in different biochores. In this connection, intermediate regions between the Boreal and Himalayan provinces, which show some relations to both biochores, are of prime importance. Among such regions, there is the Ussuri Province, including the southern Primor'e area. Data on the Lower Triassic stratigraphy in this area were used by Kiparisova and Popov (1956, 1964) in their substantiation of the Induan–Olenekian boundary.

Several Lower Triassic reference sections are known in the southern Primor'e (Burii, 1959; Korzh, 1959; Kiparisova, 1961, 1972; Zakharov, 1968, 1978, 1997). Zakharov (1994, 1996) recently proposed the section in the Tri Kamnya Cape region on the western coast of the Ussuriiskii Bay for the candidate of the Induan–Olenekian boundary stratotype.

This paper is dedicated to substantiation of a second candidate for the global Induan–Olenekian boundary stratotype (the Olenekian stage limitotype) in the southern Primor'e, namely, in the Abrek Bay region. Paleomagnetic and petrographic research was performed in the Vladivostok Petrophysical Laboratory of the Primor'e Prospecting-Survey Expedition. Magnetic susceptibility was determined with the help of KLY-2 bridge. The remanent magnetization of samples was measured using roll-generators of JR-4 type. The temperature range during thermal demagnetization was from 100 to 600°C.

INVESTIGATION HISTORY OF LOWER TRIASSIC IN SOUTHERN PRIMOR'E

The Early Triassic ammonoids from the southern Ussuri region were first described in monograph by Diener (1895). These were mainly paleontological materials collected by D.L. Ivanov. The discovery of paleontological remains in Triassic sediments of the southern Far East was of great importance for the world stratigraphy, because only three regions with complete Lower Triassic sections were known before that moment, namely, the Alps (Mojsisovics, 1882), Arctic Siberia (Mojsisovics, 1886), and the Salt Range (Waagen, 1895).

Vittenburg (1910, 1916) distinguished *"Pseudomonotis" (=Eumorphotis) iwanowi, "Terebratula" (=Fletcherithyris) margaritovi*, and *"Xenodiscus" (=Kazakhstanites) nicolai* zones in the Lower Triassic of the southern Ussuri region. He referred basal conglomerates of the section to the Jurassic, erroneously considering that the underlying flora-bearing sequence in the Novik Bay region of the Russkii Island is of Jurassic but not of the Permian age, as it was established later on.

Kiparisova (1961) suggested the following Lower Triassic zonation based, in contrast to the scheme of Vittenburg, on ammonoids representing a single invertebrate group: (1) basal layers (conglomerates); (2) *Proptychites*, originally *Meekoceras* Zone; (3) *Flemingites* Zone (sandstones, siltstones, and shales); (4) *Prosphingites* Zone (siltstones and sandstones); and (5) *Subcolumbites* Zone (sandstones, siltstones, and shales). The Induan–Olenekian boundary was originally placed at the top of the *Flemingites* Zone (Kiparisova and Popov, 1956; Burii, 1959; Korzh, 1959; Kiparisova, 1961), and afterward, when the *Flemingites* and *Prosphingites* zones were united (Kiparisova and Popov, 1964; Kiparisova, 1972), at the base of the former.

When correlating different sections in Primor'e, Zakharov (1968) was first to note a significant facies variability of the Lower Triassic sediments in the region under discussion, where it was ignored earlier. As a result, he showed that some Induan and Olenekian zonal ammonoid assemblages distinguished by Kiparisova (1961) were misinterpreted.

By the moment, the following unified Lower Triassic zonation (Zakharov, 1968, 1978, 1997; Burii *et al.*, 1972; Burii and Zharnikova, 1981; Zakharov and Rybalka, 1987) is approved for the Far East by the 4th Interdepartmental Regional Stratigraphic Conference on the Precambrian and Phanerozoic of the southern Far East and eastern Transbaikal region (Khabarovsk, 1990).

Induan Stage: (1) *Glyptophiceras ussuriense* beds; (2) *Gyronites subdharmus* Zone.

Olenekian Stage: (1) *Hedenstroemia bosphorensis* Zone; (2) *Anasibirites nevolini* Zone; (3) *Tirolites* Zone (with *Bajarunia dagysi* and *Tirolites ussuriensis* beds); (4) *Neocolumbites insignis* Zone; (5) *Subcolumbites multiformis* Zone.

Burii and colleagues (Burii *et al.*, 1976, 1977, 1993; Burii, 1997) subdivided the Induan and Olenekian stages into the Lazurnaya, Tobizin, and Chernyshev horizons. After investigation of facies variability in the Lower Triassic sequences, both stages were additionally subdivided into formations (Zakharov, 1997).

The Early Triassic conodonts from southern Primor'e were first studied by Burii (1979). It should be noted however that using conodonts it is impossible to establish the exact position of the Olenekian lower boundary anywhere in the world, as stratigraphic range of index species of the *Neospathodus pakistanensis* Zone spans the uppermost Induan and lowermost Olenekian strata in both the Salt Range and Primor'e areas.

ABREK BAY SECTION OF THE LOWER TRIASSIC

The section is situated on the northeastern coast of the Abrek Bay, 0.8 km north of the Yunshi Cape (42°55′ N, 131°26′ E; Figs. 1 and 2).

Biostratigraphic Succession

Ivanov (see in Diener, 1895) was first to detect the Early Triassic bivalve and gastropod shells in the clayey deposits of the Abrek Bay (Strelok Inlet locality). Kiparisova (1938) cited the following taxa of Lower Triassic bivalves collected by Vittenburg in 1908–1923 and Krishtofovich in 1924 around the Abrek Bay: *Claraia aurita* Hauer, *Eumorphotis multiformis* (Bittner), *"Pecten" (=Chlamys?) kryshtofowichi* Kipar., *Myoconcha* aff. *goldfussi* Dunk., and *Posidonia abrekensis* Kipar.

The first and sole description of Triassic cephalopods from the Abrek Bay region was done by Kiparisova (1961) who investigated collections by Nozdreev, Trifonov, Burii, and Korzh of 1938-56. She described two nautiloid forms (*Menuthionautilus, "Syringoceras"*), seven ammonoids [(*Hedenstroemia, Arctoceras (="Proptychites"), Gyronites*, and *Meekoceras*] from Lower Triassic beds, and a shell identified as *Discoptychites.*

Kiparisova (1972) noted a considerable discrepancy between the stratigraphic schemes of Nozdreev (Kiparisova, 1972), Burii (1959), Korzh (1959), and Vasil'ev (Kiparisova, 1972), and this eventually has led her to their invalid correlation. She determined the Induan–Olenekian boundary at the appearance level of *Hedenstroemia* and *Meekoceras*, on the one hand, and at the transition from sandy to silty facies, on the other. Her conclusion comes into conflict with report by Nozdreev who encountered *Hedenstroemia* in the sandy facies. Dagys (1974) who described a new genus and species of an articulate brachiopod *Abrekia sulcata* Dagys from

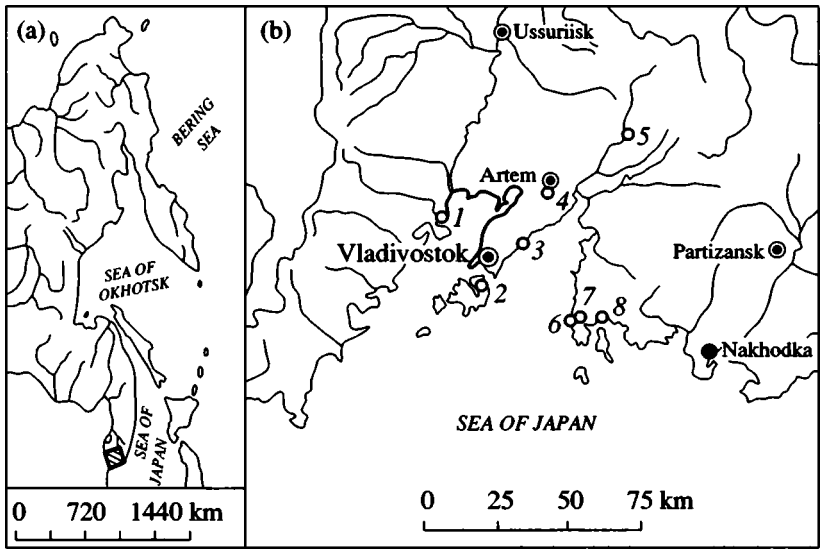

Fig. 1. Sections of the Induan–Olenekian boundary sediments (Lower Triassic) in the study area (a) crosshatched and southern Primor'e (b):
(*1*) western coast of the Amurskii Bay between the Atlasov and Ugol'nyi capes; (*2*) Ayaks Bay (Russkii Island); (*3*) Tri Kamnya Cape and Orel Cliff (western coast of the Ussuriiskii Bay); (*4*) "SMID" Quarry near the Artem; (*5*) left bank of the Artemovka River; (*6*) Kom-Pikho-Sakho Cape; (*7*) area of the Yuzhnorechensk railway station (Shimeuza); (*8*) Abrek Bay (Strelok Strait).

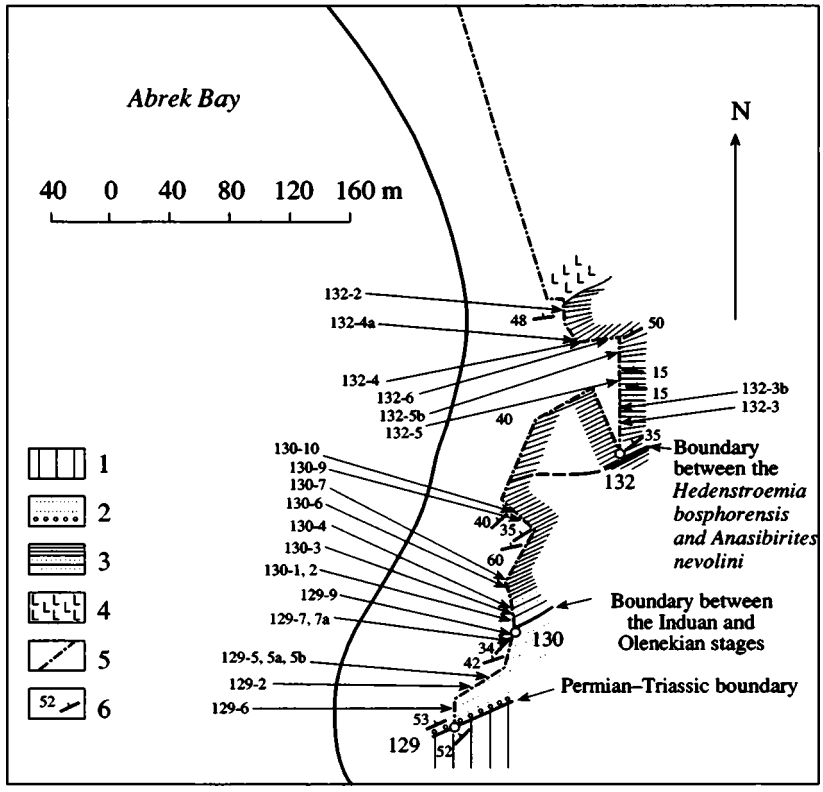

Fig. 2. Geology of the northeastern coast of the Abrek Bay: (1) continental sediments of the Murgabian Stage (Abrek Formation); (2) Induan marine sediments (lower and middle parts of the Lazurnaya Formation); (3) marine sediments of the Olenekian *Hedenstroemia bosphorensis* and *Anasibirites nevolini* zones (upper part of the Lazurnaya and lower part of the Zhitkov formations); (4) dykes of dioritic porphyrite; (5) route of investigation; (6) bed attitude.

the Lower Triassic sandstones of the Abrek Bay attributed them to the Induan Stage.

The description of the Lower and, partly, Middle Triassic sequence of the Abrek Bay region that is given below is based on more representative assemblages of invertebrates collected in 1997 and 1998. Continental deposits of the Murgabian Abrek Formation, which are represented at the top by light gray, fine- to medium-grained tuffaceous sandstones and gravelstones, are unconformably, with a hidden hiatus, overlain by the following Lower Triassic units (from the base upward).

Induan Stage

Gyronites subdharmus Zone

(1) Fine- to medium-pebbled conglomerate 1.3 m thick; pebbles are set in a greenish-gray sandy matrix differing from underlying sandstones by a good sorting. Pebbles of acidic and intermediate extrusive rocks prevail in the coarse clastic fraction.

(2) Greenish gray, fine-grained sandstone bearing numerous lenses (1 to 3 cm thick) of fine-pebbled conglomerates; total thickness is 4.0–4.5 m.

(3) Greenish gray, fine-grained sandstone enclosing separate molluscan valves, rare pebbles, small angular fragments of black siltstone and shale, and few thin lenses of fine-pebble conglomerates; the bed is 19.0 to 21.0 m thick.

(4) Greenish gray, fine-grained sandstone 8.0 m thick, bearing lenses of calcareous coquinoid sandstone up to 15 cm thick; an interlayer of banded sandstone with clayey laminae 1 to 3 mm thick occurs 5 m above the base of the member. The sediments yield remains of brachiopods *Lingula borealis* Bittner and *Orbiculoidea* sp., bivalves *Claraia australasiatica* Krumb., *Promyalina putiatinensis* (Kipar.), *Eupecten* cf. *ussuricus* (Bittner), and *Entolium microtis* Witt., ammonoid *Gyronites subdharmus* Kipar., crab nippers among arthropods, and skate teeth among chordates.

(5) Greenish gray, fine-grained, micaceous sandstone 6.0 m thick, enclosing numerous black siltstone and shale interbeds of 2 to 10, rarely to 40-mm in thickness, and calcareous coquinoid sandstone lenses 20 to 40 cm thick; asymmetric ripple marks on bedding planes represent traces of a southeastward (125°) current. Bivalve *Neoschizodus laevigatus* (Zieten), ammonoid *Lytophiceras?* sp., and large gymnosperm leaves were found in the member.

(6) Greenish gray, fine-grained sandstone 5.0 m thick, with thin black siltstone and shale interbeds separated by intervals of 0.5–0.6-m and with a calcareous coquinoid sandstone bed 30 cm thick at the base of the member; The bedding planes are decorated with asymmetric ripple marks. Sediments contain brachiopods *Lingula* sp. and *Orbiculoidea* sp., bivalves *Eumorphotis multiformis* (Bittner), *Neoschizodus laevigatus* (Zieten), and *Anodontophora fassaensis* (Wissm.), gastropods, and ammonoids *Gyronites subdharmus* Kipar., *Koninckites* sp. indet., and *Ambites?* sp.

The total thickness of Induan sediments in the section is 43 to 46 m.

Olenekian Stage

Hedenstroemia bosphorensis Zone

(7) Gray, fine-grained sandstone 2.6 m thick, enclosing a gray calcareous coquinoid sandstone interbed (30 cm) at the base, rare thin black siltstone and shale interlayers, and siderite concretions; sediments yield bivalve mollusks *Promyalina* sp., *Entolium microtis* (Witten.), *Velopecten minimus* Kipar., Pectinidae gen. et sp. nov., ammonoids *Meekoceras boreale* Diener, and *Ambites* sp., and nautiloid *Phaedrysmocheilus* sp.

(8) Light gray, fine-grained sandstone 1.8 m thick, intercalated with fine-grained, banded clayey sandstones and calcareous coquinoid sandstone lenses up to 50 cm thick; asymmetric ripple marks are seen on bedding planes. Brachiopods *Lingula borealis* Bittner, *Orbiculoidea* sp., and dominant *Abrekia sulcata* Dagys, scarce bivalves *Promyalina* sp., ammonoids *Arctoceras?* sp. indet., *Meekoceras boreale* Diener, and Melagathiceratidae gen. et sp. nov. are found along with few bryozoan remains.

(9) Alternating gray fine-grained and banded sandstones 2.15 m thick, enclosing numerous angular black siltstone fragments and a black silty shale layer 5 cm thick at the base; ammonoids Melagathiceratidae gen. et sp. nov. were encountered in sediments.

(10) Alternating light gray, fine-grained, banded (80 cm thick) and greenish gray fine-grained (5 cm thick) sandstones bearing scarce calcareous marly concretions; the total thickness is 8.0 m. The member yields nautiloids *Menuthionautilus korzchi* Kiparisova (Shigeta's collection) and *Gyronautilus praevolutum* (Kiparisova) (Zakharov and Shigeta, 2000), and ammonoids *Meekoceras boreale* Diener. The described ammonoid *Hedenstroemia* sp. indet. (Nozdreev's collection) and nautiloid *Menuthionautilus korzchi* Kipar. (Korzh's collection) (Kiparisova, 1961, 1972) most likely originated from these sandstones.

(11) Alternating black siltstone and argillite beds 50 cm thick, greenish-gray fine-grained banded sandstone layer up to 20 cm thick, and gray fine-grained sandstone bed 5 to 10 cm thick; total thickness is 2.7 m.

(12) Greenish gray, fine-grained sandstone, banded owing to thin shale laminae and bearing calcareous marly lenses and concretions; thickness is 10.0 m. Sediments contain bivalve mollusks *Palaeoneilo? prynadai* Kipar., *Pteria ussurica* Kipar., *Eumorphotis iwanowi* (Bittner), *Promyalina* sp., and *Anodontophora fassaensis* (Wissm.) and ammonoids *Hedenstroemia bosphorensis* (Zakharov) (Shigeta's collection), *Parahedenstroemia conspicienda* Zakharov, *Inyoites spicini* Zakharov, *Arctoceras septentrionale* (Diener), *Pseudoprosphingites magnumbilicatus* (Kipar.),

Ambites sp., *Koninckites* aff. *timorensis* Wanner, *Meekoceras boreale* Diener, *M. varaha* Diener, *Dienerоceras chaoi* Kipar., and *Preflorianites* cf. *radiatus* Chao.

(13) Greenish gray siltstone 8.6 m thick, banded because of thin shale laminae and bearing scarce calcareous marly concretions; ammonoids *Meekoceras varaha* Diener and *Meekoceras* sp. nov. were encountered at this level.

(14) Black siltstone 23.0 m thick, enclosing black shale beds with calcareous marly concretions and lenses, and rare gray fine-grained sandstone interlayers up to 15 cm thick; near the base, the member yields small bivalves *Velopecten minimus* Kipar., small gastropods, ammonoids *Pseudoprosphingites magnumbilicatus* (Kipar.), Gyronitidae?, *Koninckites timorensis* Wanner, dominant *Meekoceras varaha* Diener, *Flemingites radiatus* Waagen, and *Anaxenaspis* cf. *orientalis* (Diener).

Anasibirites nevolini Zone

(15) Alternating black siltstone and shale beds 35.0 m thick in total, bearing large calcareous marly lenses and concretions; sediments yield brachiopods *Abrekia?* sp., small bivalves *Promyalina* sp. and *Posidonia?* sp., and ammonoids *Pseudosageceras longilobatum* Kipar., *P.* sp., dominant *Arctoceras labogense* (Zharnikova), *A. subhydaspis* (Kipar.), *A. septentrionale* (Diener), *Pseudoprosphingites magnumbilicatus* (Kipar.), *Owenites koeneni* Hyatt et Smith, Gyronitidae gen. et sp. nov., dominant *Koninckites timorensis* Wanner, *Gurleyites* sp., *Anasibirites nevolini* Burij et Zharnikova, *Palaeokazakhstanites ussuriensis* (Zakh.), *Euflemingites prynadai* (Kipar.), and *Eophyllites* sp.

(16) Black siltstone and shale beds (5.0 m thick in total) with banded siltstone interlayers and calcareous marly concretions; ammonoids *Arctoceras labogense* (Zharnikova), *Pseudoprosphingites magnumbilicatus* (Kipar.), *Koninckites timorensis* Wanner, *Parakymatites* sp. nov., *Meekoceras varaha* Diener, *M.* sp. nov., *Hemiprionites dunajensis* Zakh., and *Preflorianites* cf. *radiatus* Chao. are found in the member. *Arctoceras abrekense* (Kipar.) encountered by Nozdreev, is likely derived from shales of the member.

(17) Gray banded siltstone 2 to 3 m thick.

The total thickness of two lower Olenekian zones is about 100 m.

Analysis of Faunal Assemblages

Bryozoans. Few fragments of branching bryozoan colonies were found in the calcareous sandstones of the lower Olenekian.

Brachiopods. About five or six brachiopod species are known from the Lower and Middle Triassic of the Abrek Bay region (Zakharov and Popov, 1999); all forms known from Induan sediments are inarticulate (in fauna). The lower Olenekian deposits are characterized by appearance and mass development of articulate brachiopod taxon *Abrekia sulcata.* Inarticulate *Lingula borealis* and *Orbiculoidea* sp. are typical of the Induan and lower Olenekian sandstones. *Orbiculoidea* forms are dominant in the upper Induan.

Bivalve mollusks. Bivalves, like other faunal remains, are extremely rare in the lower Induan sediments represented by shallow-water coarsely clastic facies, which contain only separate valves or broken shells. Similar bivalve assemblages are recorded in the upper Induan and lower Olenekian sandstones. Common species from these facies are *Entolium microtis, Neoschizodus laevigatus, Anodontophora fassaensis,* and some others. Among species abundant in the Induan sediments only, we can mention *Promyalina shamarae* and, likely, *P. putiatinensis,* though *Promyalina* sp. similar in morphology to the latter, occurs in the Olenekian as well.

Nautiloids. Remains of Early Triassic nautiloids were encountered only in the Olenekian sediments of the Abrek Bay. The trochospiral Olenekian forms are represented by species *Phraedrysmocheilus* sp., *Menuthionautilus korzhi,* and *Gyronautilus praevolutum.*

Ammonoids. Among Induan ammonoids only four species were classified. These are the zonal index species *Gyronites subdharmus, Lythophiceras?* sp., *Koninckites* sp. indet., and *Ambites?* sp. The *Hedenstroemia bosphorensis* Zone, the basal one in the lower Olenekian, is defined above the first occurrence level of *Meekoceras* forms, which are associated higher with *Hedenstroemia* and some other typical Olenekian taxa. The ammonoid assemblage of the zone is represented by 19 species of 14 genera (Fig. 3). The zonal index species was recently found in the lower part of the unit (Bed 12).

The base of overlying *Anasibirites nevolini* Zone marks the appearance level of its index species. *Arctoceras labogense* (dominating), *A. subhydaspis* (Kipar.), *A. abrekense, Gurleyites* sp., *Palaeokazakhstanites ussuriensis, Hemiprionites dunajensis, Eophyllites* sp., and *Parakymatites* sp. nov. are characteristic of the zone as well. Many other species, namely, *Arctoceras septentrionale, Pseudoprosphingites magnumbilicatus, Owenites koeneni, Koninckites timorensis, Meekoceras varaha, Euflemingites prynadai,* and others, are also known from the underlying zone of the described and other regional sections. In total, 20 species of 15 genera represent the assemblage of the *Anasibirites nevolini* Zone of the Abrek Bay section. Among them, one genus and two species are new.

Arthropods. Nipper remains of small crabs were encountered in the Induan sandstones.

Chordates. Presence of amphibian remains in the Olenekian clayey sediments (*Anasibirites nevolini* Zone) is inferred from found small fragments of bony

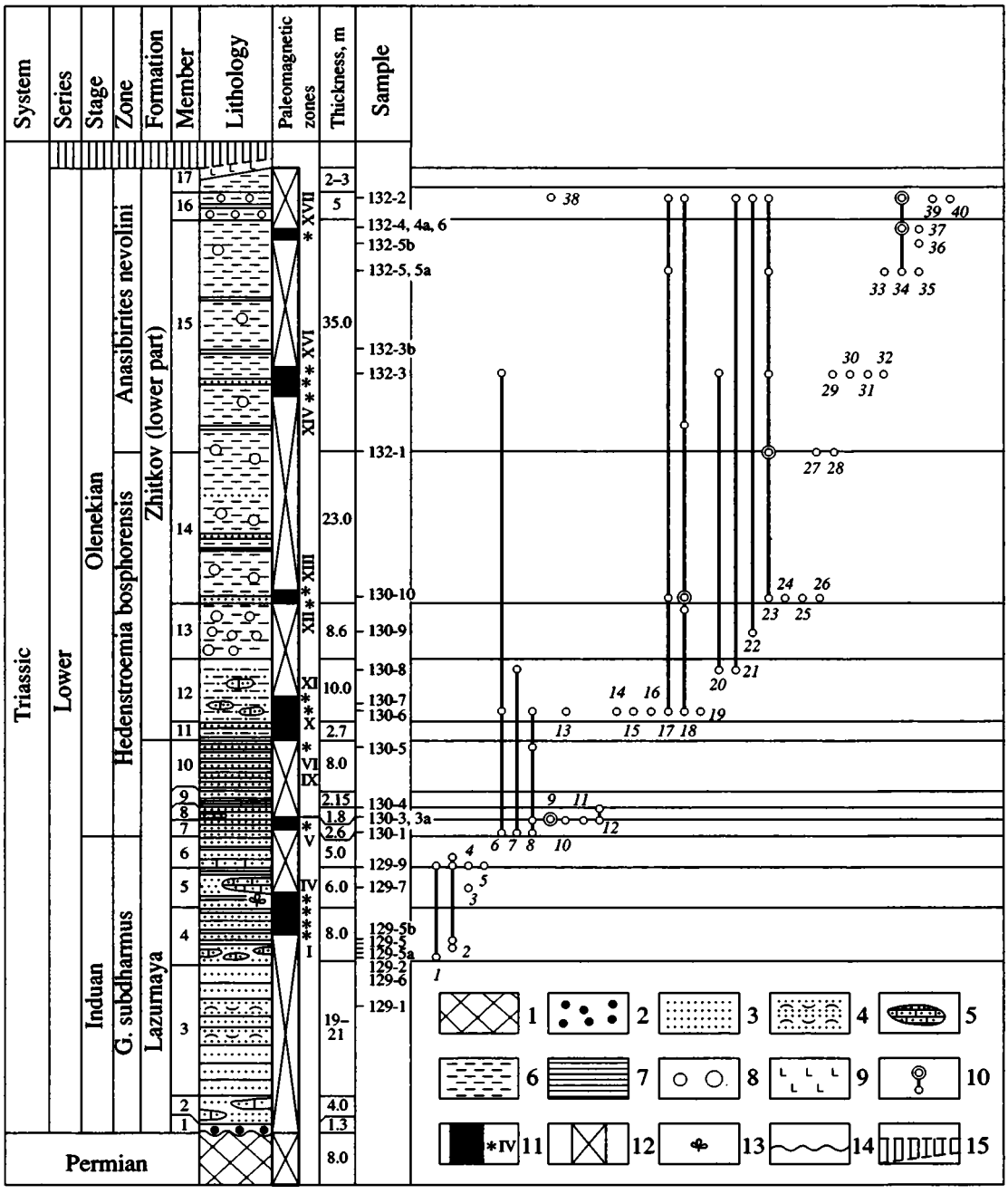

Fig. 3. Major invertebrate groups in the Lower and Middle Triassic of the Abrek Bay. Lithology: (1) Permian (Murgabian) sediments of the Abrek Formation; (2) conglomerate; (3) fine-grained sandstone; (4) sandstone bearing separate molluscan valves; (5) calcareous coquinoid sandstones; (6) silty shale; (7) shale; (8) calcareous marly concretions; (9) porphyrite dykes; (10) species range (dominating levels are marked by double circles); (11) normally magnetized rocks (asterisks and Roman numerals mark levels sampled for paleomagnetic measurements; (12) intervals unstudied for paleomagnetic properties; (13) leaf flora; (14) unconformity; (15) hidden interval. Species: (*1*) *Gyronites subdharmus* Kiparisova; (*2*) *Promyalina putiatinensis* (Kiparisova); (*3*) *Lytophiceras?* sp.; (*4*) *Promyalina schamarae* (Bittner); (*5*) *Koninckites* sp. indet.; (*6*) *Promyalina* sp.; (*7*) *Ambites* sp.; (*8*) *Meekoceras boreale* Diener; (*9*) *Abrekia sulcata* Dagys; (*10*) *Hedenstroemia* sp. indet.; (*11*) *Arctoceras?* sp. indet.; (*12*) Melagathiceratidae gen. et sp. nov.; (*13*) *Hedenstroemia bosphorensis* (Zakharov); (*14*) *Inyoites spicini* Zakh.; (*15*) *Koninckites* aff. *timorensis* Wanner; (*16*) *Dieneroceras chaoi* Kiparisova; (*17*) *Pseudoprosphingites magnumbilicatus* (Kiparisova); (*18*) *Meekoceras varaha* Diener; (*19*) *Parahedenstroemia conspicienda* Zakharov; (*20*) *Arctoceras septentrionale* (Diener); (*21*) *Preflorianites* cf. *radiatus* Chao; (*22*) *Meekoceras* sp. nov.; (*23*) *Koninckites timorensis* Wanner; (*24*) Gyronitidae?; (*25*) *Flemingites radiatus* Waagen; (*26*) *Anaxenaspis* cf. *orientalis* (Diener); (*27*) *Gurleyites* sp.; (*28*) *Anasibirites nevolini* Burij et Zharnikova; (*29*) *Owenites koeneni* Hyatt et Smith; (*30*) Gyronitidae gen. et sp. nov.; (*31*) *Palaeokazakhstanites ussuriensis* (Zakharov); (*32*) *Eophyllites* sp.; (*33*) *Pseudosageceras* sp. indet.; (*34*) *Arctoceras labogense* (Zharnikova); (*35*) *Euflemingites prynadai* (Kiparisova); (*36*) *Arctoceras subhydaspis* (Kiparisova); (*37*) *Pseudosageceras longilobatum* Kiparisova; (*38*) *Abrekia?* sp.; (*39*) *Parakymatites* sp. nov.; (*40*) *Hemiprionites dunajensis* Zakharov.

Table 1. Physical properties of the Lower Triassic sediments from the Abrek Bay region

Stratigraphic level	Mean parameter values			
	n	σ	χ	*In*
Induan (*Gyronites subdharmus* Zone)	49	2.550	186	0.8
Olenekian (*Hedenstroemia bosphorensis* and *Anasibirites nevolini* zones)	144	2.632	197	8.2

Note: (*n*) number of measurements; (σ) density (g/cm^3); (χ) magnetic susceptibility (10^{-6} SI); (*In*) remanent magnetization (mA/m).

Table 2. Paleomagnetic characteristics of Triassic sediments from the Abrek Bay locality

Stage	*N*(*n*)	*D*	*J*	*k*	*a*	Lt	Lng	O1	O2	Ltm	DMCM
Olenekian	10(23)	313.5	43.2	26.5	9.6	48.2	32.7	12	7	25.2	T400
Induan	4(9)	303.6	30.3	60.5	16.0	35.5	31.4	18	10	16.3	T400

Note: *N*(*n*), number of samples; (*D*, *J*) declination and inclination of the average vector of characteristic remanent magnetization; (*k*) concentration parameter of vectors; (*a*) confidence circle radius for the average vector at probability of 0.95; (Lt, Lng (O1, O2)) latitude and longitude (half axes of the definition error ellipsis) of the paleomagnetic north pole; (Ltm) geomagnetic latitude (paleolatitude) of the region; (DMCM) definition method of characteristic magnetization; (*T*) temperature of thermal demagnetization.

tissue. Skate teeth are found in the Induan and Olenekian sandstones.

Paleomagnetic Characteristics of the Induan–Olenekian Boundary Beds

As is known (Zakharov and Sokarev, 1991), it is difficult to use paleomagnetic data for the purpose of Lower Triassic magnetostratigraphy in southern Primor'e, because the investigated Triassic rocks from the Russkii Island and western coast of the Ussuriiskii Bay presumably reveal the metachronous (apart from primary and viscous) remanent magnetization of unclear origin. At the same time, the latest preliminary study showed that the Triassic rocks of the Abrek Bay locality are well suitable for magnetostratigraphic research. Sedimentary rocks, which were sampled here in a rather loose manner, revealed normal magnetization at 17 stratigraphic levels of the upper *Gyronites subdharmus* (Samples I–IV), *Hedenstroemia bosphorensis* (V–XIII), and *Anasibirites nevolini* (XIV–XVII) zones of the Induan and Olenekian stages (Fig. 3). At least some of them likely characterize a single extended zone of normal magnetization.

Table 1 includes data on physical properties of 17 orientated Lower Triassic rocks samples from the southeastern coast of the Abrek Bay. As one can see, the Induan rocks have lower density and remanent magnetization as compared to the Olenekian sediments despite the comparable parameters of magnetic susceptibility at both levels. The correlation analysis underlines these distinctions: a stable direct correlation between density and magnetic susceptibility is characteristic of the Induan rocks only. At the same time, rocks of both stages exhibit similar relationships between magnetic susceptibility and remanent magnetization parameters.

During the stepwise demagnetization by sample heating to 500°C, magnetic susceptibility of rocks remains almost constant, and changes do not exceed 10% of the initial value. As the heating temperature raises, the remanent magnetization of samples decreases down to 10–40% of the initial value at 400°C and to several percent at 500°C. At 600°C, the mineral composition changes considerably whereas magnetic susceptibility and remanent magnetization increase two–three times and more. Accordingly, we included in Table 2 only those paleomagnetic characteristics of studied Lower Triassic sediments, which have been measured by the thermal demagnetization up to the temperature of 400°C.

The presented data on the Abrek Bay section and published results on other Lower Triassic sections of southern Primor'e (Zakharov, 1994, 1996) show that beds, which yield the early Olenekian ammonoid assemblage, are easily distinguishable in the region. As far as it concerns the definition of the Induan–Olenekian boundary, the main problem is that the boundary beds proper are insufficiently exposed in the most known sections.

LOWER TRIASSIC LITHOLOGIC FACIES AND REGIONAL CORRELATION IN SOUTHERN PRIMOR'E

There are two major types of Lower Triassic sequences in southern Primor'e, which are composed of many and two lithologic facies. Western sections, which are situated in the Russkii Island and along western coast of the Amurskii Bay between the Atlasov and Ugol'nyi capes, are of the first type. Eastern sections of the second type are located in the following areas: the left bank of the Artemovka River, the Artem region, the Kom-Pikho-Sakho Cape on the eastern coast of the

Table 3. Correlated ammonoid and conodont zonations in the upper Induan and lower Olenekian sediments

Stage	Substage	Southern Primor'e: Zakharov, 1997	Southern Primor'e: Burii, 1979 (modified)	Salt Range: Waagen, 1895; Guex, 1978 (modified)	Salt Range: Sweet, 1970	Northeastern Asia: Dagys, Ermakova, 1993 (modified)	Northeastern Asia: Dagys, 1984	Canada: Orchard, Tozer, 1997	
Olenekian (lower part)	Ayaxian	*Tirolites–Amphistephanites* / *Tirolites ussuriensis*	*Icriospathodus collinsoni*	*Tirolites–Columbites* (lower part)	*Neogondolella jubata* (lower part)	*Northophiceras contrarium*	*Neogondolella jubata* (lower part)	?	*Neogondolella* aff. *sweeti*
		Tirolites–Amphistephanites / *Bajarunia dagysi*				*Bajarunia euomphala*			
		Anasibirites nevolini	*Neospathodus waageni* / *Neogondolella milleri*	*Anasibirites pluriformis*	*Neospathodus waageni*	*Anawasatchites tardus*	*Neospathodus waageni* / *Neogondolella milleri*	*Anawasatchites tardus*	*Neogondolella milleri*
		Hedenstroemia bosphorensis	*Neospathodus waageni* / *Parachirognathus–Furnishius*	*Flemengites flemengianus*	*Neospathodus waageni*	*Lepiskites kolymensis* / *Hedenstroemia hedenstroemi*	*Neospathodus waageni* / *Neogondolella moscheri*	*Euflemingites romunderi*	*Gladigondolella meeki* / *Neospathodus pakistanensis*
		Hedenstroemia bosphorensis	*Neospathodus pakistanensis*		*Neospathodus pakistanensis*		*Neogondolella nepalensis*	*Hedenstroemia hedenstroemi*	?
Induan (upper part)		*Gyronites subdharmus* (upper part)	*Neogondolella carinata* (upper part)	*Prionolobus rotundatus* (upper part)	*Neospathodus cristagalli*	? / *Tompoprotychites turgidus* / *Vavilovites sverdrupi*	?	*Vavilovites sverdrupi*	*Neospathodus cristagalli*

Ussuriiskii Bay, vicinity of the Yuzhnorechensk railway station, and the Abrek Bay region. The Tri Kamnya Cape section on the western coast of the Ussuriiskii Bay is the intermediate between those groups, though closer to the western one.

The Russkii Island is a standard area of the polyfacies type sections. Zakharov (1997) distinguished here the following subdivisions: (1) the Lazurnaya Formation in the range of the Induan Stage and lower beds of the Olenekian *Hedenstroemia bosphorensis* Zone, which are most completely exposed in the Ayaks Bay of the Russkii Island, though the formation stratotype is situated in the western coast of the Ussuriiskii Bay between the Lazurnaya Inlet and the Tri Kamnya Cape (Zakharov, 1968); (2) the Tobizin Formation in the range of the *Hedenstroemia bosphorensis* Zone, its basal beds excluded, and the *Anasibirites nevolini* Zone, stratotype of which is located in the Tobizin Cape, Russkii Island; (3) the Schmidt Formation in the range of the *Tirolites-Amphistephanites* Zone with the stratotype on the Schmidt Cape, Russkii Island; and (4) the Zhitkov Formation in the range of the *Neocolumbites insignis* and *Subcolumbites multiformis* zones with the stratotype in the Zhitkov Cape, Russkii Island. Formations one to three are mainly represented by a sandy facies, though they differ from each other. The Lazurnaya Formation (110 m thick) shows predominance of coarse rocks in the lower part and occurrence of bivalve coquina lenses in the middle and upper parts. The Tobizin Formation (180 m thick) exhibits presence of cephalopod coquina beds and the appearance of a thin silty shale bed. The Schmidt Formation (40 m thick) encloses numerous and rather thick (up to 1.2 m) bivalve and brachiopod coquina lenses. The Zhitkov Formation (85 m thick) differs sharply from the underlying Lower Triassic formations in a wide distribution of silty shales bearing abundant ammonoid and bivalve remains.

In the eastern group of sections, the Lower Triassic is represented by two subdivisions only, the Lazurnaya and Zhitkov formations. The distinct boundary between them is within the lower Olenekian *Hedenstroemia bosphorensis* Zone, as it is exemplified by the section in the Abrek Bay southeastern coast (see Fig. 3).

In the eastern group of sections of southern Primor'e, the Zhitkov Formation corresponds to the *Hedenstroemia bosphorensis* (its lower part excluded), *Anasibirites nevolini, Tirolites-Amphistephanites, Neocolumbites insignis* and *Subcolumbites multiformis* zones. Only the lower part of the unit corresponding to the Tobizin Formation of the Russkii Island is exposed at the Abrek Bay locality.

The lower boundary of the Zhitkov Formation in southern Primor'e is obviously diachronous. The consecutive facies changes represent a result of the general basin deepening that progressed slower in its western or southwestern parts, as it is evident from lithology of the western sections.

CONCLUSION

The comprehensive analysis indicates that only two Lower Triassic sections of southern Primor'e, namely, that in the Tri Kamnya Cape area of the Ussuriiskii Bay western coast and the Abrek Bay section near the Strelok Inlet may be suggested as candidates for the Global Stratotype Section and Point of the Induan–Olenekian boundary. All alternative sections in southern Primor'e, despite some their disadvantages, may elucidate the total composition of the upper Induan–lower Olenekian zonal assemblages thus contributing to solution of the global correlation problem.

The Olenekian Stage base in southern Primor'e, Himalayas, Siberia, and Canada can be defined at the appearance level of *Hedenstroemia* forms coexisting with *Meekoceras* forms in the Tethyan Province. The recently discovered association of *Meekoceras gracilitatis* White with *Flemingites* and other ammonoid taxa typical of the *Hedenstroemia bosphorensis* Zone (Zakharov's collection from the western coast of the Ussuriiskii Bay in southern Primor'e) suggests that this zone is reliably correlative with the *Meekoceras gracilitatis* Zone of Idaho and with the *Flemingites flemingianus* Zone of the Salt Range. Currently it has been also found in southern Primor'e that *Euflemingites prynadai* (Kipar.) and *Arctoceras* forms are widespread in two lower zones of the Olenekian Stage. Since *Euflemingites* and *Arctoceras* species are also known in the Boreal areas, they may be used to refine correlation between the Smithian regional stage of Canada (Tozer, 1994) and lower Olenekian zones of the Primor'e (the regional Ayaxian Substage, Zakharov, 1997).

Judging from preliminary magnetostratigraphic data and distribution of *Arctoceras* and *Euflemingites* species, the inferred interval of an extended normal-polarity zone of the Ayaxian Substage in southern Primor'e can be correlated with the normal-polarity zone of comparable range that was recently established in Spitsbergen and Canada. In the last case, it spans at least the *Euflemingites romunderi* (upper part) and *Wasatchites tardus* zones of the Smithian regional stage (Mork *et al.*, 1999).

Dagys and Ermakova (1993); Dagys (1997) correctly noted the difficulty of correlating the Induan–Olenekian boundary sediments, when they are situated in different biogeographic areas despite an excellent global correspondence of the *Anasibirites* and *Wasatchites* (*Neogondolella milleri* Zone) beds (Table 3) and of some other Lower Triassic units of the conodont zonation. This is quite natural, though the problem is complicated, from our standpoint, by some fallacies in the Lower Triassic biostratigraphic scheme of northeastern Asia (Dagys and Ermakova, 1993), which has been repeatedly under debates (Zakharov, 1994a, 1994b, 1995, 1996; Dagys, 1995, 1997). It is appropriate to mention here just two points.

Paleontological characteristics of upper Induan deposits in northeastern Asia are inadequately investi-

gated because of an extreme rarity of ammonoids and conodonts at this stratigraphic level. This explains to some extent so frequent revisions of the upper Induan biozonation in the region without any significant contribution of new factual data. In the new Lower Triassic biostratigraphic scheme of northeastern Asia, the *Kingites? korostelevi* Zone (Dagys and Ermakova, 1993) distinguished in the Burgagandzha River section of the eastern Verkhoyansk region was unsuccessfully suggested as an upper unit of the stage. The zone is obviously invalid, because all 16 investigated specimens of *Kingites korostelevi* Zakharov are obtained from the lower (Zakharov 1978), but not upper Induan strata in interpretation by Dagys and Ermakova (1993). Moreover, these ammonoids were found by Zakharov in association with the early Induan *Episageceras, Tompophiceras*, and *Ophiceras* forms.

Another innovation in zonation under discussion is the Olenekian *Lepiskites kolymensis* Zone placed by the aforementioned authors immediately above the *Hedenstroemia hedenstroemi* Zone. This contradicts the available records in the Ken'elichi River section of the Kulu River basin (Zakharov, 1978), where the *Lepiskites kolymensis* (Popow) holotype was found. The upper Induan sediments of this section yield mollusks *Promyalina schamarae* (Bittner), *Vavilovites (Vavilovites) kuluensis* Zakharov, and *Prionolobus* sp. nov. in association with abundant conchostracans *Lioesteria* (Zakharov and Vavilov, 1976; Zakharov, 1978). The early Olenekian ammonoids of the section are concentrated in large concretions at stratigraphic levels of 46 and 60 m above the base of the Olenekian Stage. The ammonoids are represented here mainly by *Hedenstroemia mojsisovicsi* Popow (= *H. borealis* Popow, *H. serta* Popow, *"Anahedenstroemia" tscherskii* Popow) typical of the *Hedenstroemia hedenstroemi* Zone. In some concretions from the mentioned levels, one of us encountered *H. mojsisovicsi* in association with *Lepiskites kolymensis* (Popow), *Pseudosageceras multilobatum* Noetling, and *Sakhaites subleptodiscus* (Popow). In the overlying beds up to 220 m thick, *Hedenstroemia mojsisovicsi* coexists with *Lepiskites* sp. and *Sakhaites subleptodiscus* (Popow) (Zakharov, 1978). The *Hedenstroemia* beds of the Kulu River basin are overlain by sediments yielding *Wasatchites* cf. *meeki* Mathews of the *Wasatchites tardus* Zone (Bychkov, 1972). Therefore, ranges of *Hedenstroemia* and *Lepiskites* forms most likely coincide in the upper reaches of the Kolyma River, and occurrence of *Lepiskites kolymensis* in the lower 50 to 60-m thick member of the Olenekian deposits is reliably established here. This fact casts doubt on validity of the *Lepiskites kolymensis* Zone distinguished in some regions of northeastern Asia.

ACKNOWLEDGMENTS

We are grateful to V.A. Zakharov, T.B. Leonova, V.R. Lozovskii, P.V. Markevich, and A.A. Shevyrev for their critical remarks and to L.I. Sokur for technical assistance in preparation of the manuscript. The work was supported by the Russian Foundation for Basic Research, project nos. 97-05-65832, 98-05-65343, and 98-04-49355.

Reviewers T.B. Leonova and A.A. Shevyrev

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Stratigraphy and Geological Correlation, Vol. 10, No. 6, 2002, pp. 586–602. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 10, No. 6, 2002, pp. 62–81.
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Climate during Permian–Triassic Biosphere Reorganizations, Article 1: Climate of the Early Permian

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Received September 19, 2000

Abstract—The Permian was marked by transition from the glacioera to the thermoera, the last one in the geological history. This global climatic reorganization followed the significant paleogeographic changes related to Pangea formation and preceded the major reorganization in the Earth's biota that occurred in the terminal Permian–initial Triassic. Climatic changes served as a coupling process in this chain of events. The analysis of climatic history makes it possible to estimate the role of climate in the cardinal reorganization of the Late Paleozoic biosphere. The compiled one lithological–paleogeographic and two paleoclimatic maps allowed reconstruction of the Early Permian climatic zonation. In the initial Permian, glaciations developed in high and middle latitudes of the Southern Hemisphere and resembled the Pleistocene glacial maximum. Later on, the ice shield gradually decreased and in the late Sakmarian–initial Artinskian, it approached the shape of a polar cap. Separate glaciation centers were preserved only in Antarctica until the end of the Permian. These climatic changes were not unidirectional and showed different-scale oscillations. The Permian climate was strongly influenced by the Pangea supercontinent. The asymmetric position of continental blocks relative to the equator determined by an asymmetry in climatic zoning. It was particularly well manifested at the beginning of the Permian, when the glacial belt in the past continental southern hemisphere extended from the pole to 40°–30°S. Correspondingly, other climatic belts in the southern hemisphere were reduced except for the arid one. To the contrary, the glacial belt was missing in the oceanic northern hemisphere, where most developed were semiarid and temperate belts. By the end of the Early Permian, the asymmetry in position of climatic belts slightly decreased. Arid and semiarid climatic belts were widely developed in both hemispheres. The growing aridity of the Pangea supercontinent was caused by its huge dimension, orographic barriers, and successive regression of epicontinental seas and strengthened by the global postglacial warming.

Key words: Early Permian, glacial, cool–temperate, temperate, semiarid, arid, and equatorial climatic belts, climatic asymmetry, glaciations, oscillations.

INTRODUCTION

The Permian–Triassic interval of the geological history was characterized by three peculiar features, which make it an attractive object for the study of biosphere reorganizations. First, the geographic reorganization that resulted in the formation of principally new configuration of oceans and single Gondwana supercontinent stretched from the South Pole to almost the Northern Pole terminated by its beginning. Second, this interval was marked by major climatic reorganizations: by transition from the Gondwana glacial era (glacioera) to the Siberian ice-free era (thermoera) and by successive aridity growth in Pangea due to its huge dimension, occurrence of mountains isolating it from the ocean, regression of epicontinental seas, and general planetary warming.[1] Third, the largest Phanerozoic biotic crisis that started in the mid-Late Permian, was particularly strong near the Permian–Triassic boundary and resulted in the replacement of the "Paleozoic evolutionary fauna" by the Recent one (Sepkowski, 1990).

Thus, the geographic reorganization was followed by climatic and biotic ones. Their integral consequences were very significant. They imply that the entire biosphere was completely transformed during the Permian–Early Triassic from the "cool" (glacial) to the "warm" (ice-free) one. The climate served as a coupling agent in this reorganization. The analyzed dynamics climatic of history, first of all, the revealed climatic zoning makes it possible to estimate the role of climate in this biosphere reorganization and to elucidate relationships between its main events.

In addition, the Permian–Triassic is the closest interval of geological history most suitable for investigations of transition from the glacial climate of the "cool biosphere" to the ice-free climate of the "warm biosphere." The study of climatic changes during the Permian–Early Triassic and their consequences have therefore a great scientific and prognostic significance for the assessment of similar changes, which can occur in the future. The preceding, older transition from the

[1] This ice-free era is frequently termed the "Mesozoic Era," but this is not quite correct because it spanned, in addition to the Mesozoic, the early Cenozoic Era.

cool to warm state of biosphere occurred in the terminal Ordovician, and it is a substantially more complicated task to study this phenomenon because of many reasons.

Paleogeographic, biotic, partly sedimentological, and climatic events of the Permian have recently been discussed in several summarizing works (Embry *et al.*, 1994; Deynoux *et al.*, 1994; Scholle *et al.*, 1995; Martini, 1997; Ziegler *et al.*, 1998; Rees *et al.*, 1999; Zharkov and Chumakov, 2001). In this paper therefore, we pay particular attention to evolution of the climatic zoning, which has not been considered at all or is characterized very briefly in these works. Our interpretations are based on paleoclimatic reconstructions that reflect typical intervals of the Permian–initial Triassic climatic history. They were compiled using the technique described earlier (Chumakov *et al.*, 1995), maps of sedimentation settings (Zharkov and Chumakov, 2001), paleogeographic data (Dobruskina, 1982, 1994; Meyen, 1987; Durante, 1995; Grunt, 1995; Ignat'ev and Naugol'nykh, 2001; Grunt and Shi, 1997; Wnuk, 1996; Shi and Grunt, 2000), and aforementioned and other paleoclimatic, paleobiogeographic, and geological materials available in publications. Difficulties in the global correlation of Permian and Lower Triassic deposits are well known. We used correlation schemes proposed for the International project "Pangea" (Ross and Ross, 1995). There are several other schemes slightly different from the latter (Kotlyar, 1997; Jin *et al.*, 1997; Kozur, 1998; Leonova, 1999; and others). Existing uncertainties in correlation determine conditional character of age interpretations of proposed paleoclimatic reconstructions and considered events, though do not influence their essence and succession.

In this paper, we discuss climatic zoning and climatic history of the Early Permian Epoch, while the next one will be dedicated to the same aspects of the Late Permian–Early Triassic (Induan) and to general problems of climatic evolution during the Permian–initial Triassic, its causes, and probable consequences.

CLIMATIC ZONING IN THE EARLY PERMIAN

The Early Permian climatic zoning was largely inherited from the Late Carboniferous one, but the trend of climatic changes was opposite in many respects. The tendency of cooling was gradually replaced by warming, and aridity of Pangea increased by the end of the Permian. Both processes cardinally changed the climatic zoning of the planet.

To reconstruct the Early Permian climatic zoning, we compiled, in addition to previously published lithological–paleogeographic map for the late Sakmarian–early Artinskian time (Zharkov and Chumakov, 2001), a similar map for the Asselian Age (Fig. 1) and two paleoclimatic maps for the late Asselian–early Sakmarian and late Sakmarian–early Asselian intervals (Figs. 2, 3). When compiling them, we used materials published in a series of works (Bhattacharya, 1991; Collinson and Miller, 1991; Cole and McLachlan, 1991; Geslin, 1998; Heckel, 1980; Isbell and Collinson, 1991; Peterson, 1980; Rall, 1996; Stevens, 1991a, 1991b; Tsubone *et al.*, 1991; Wartiti *et al.*, 1990; Wopfner, 1991), in addition to others mentioned in the text. The lithological–paleogeographic map demonstrates the sedimentation zoning at the moment of widest development of glaciation and complete formation of the collision orogenic system in central Pangea.

Below, we consider successively climatic belts of both hemispheres from high to low latitudes.

The southern glacial belt of high and middle latitudes. Glaciations of southern Pangea represent most remarkable climatic events of the initial Permian. They were responsible for formation of the sublatitudinal glacial belt (Fig. 1—SG; Figs. 2, 3—G). The development history of this belt is intricate. Most complete glacial sections discussed below demonstrate repeated alternation of continental and glaciomarine sediments with interglacial deposits. This implies a succession of different-rank glacial and interglacial phases and relevant episodes of the belt expansion and reduction accompanied by glacioeustatic sea-level fluctuations and glacioisostatic changes in the continent topography. Even dating and correlation of major (glacial and interglacial) periods is sometimes very difficult because of a limited amount of organic remains in glacial sections, endemic character of fauna and flora, and great facies variability of sediments. This explains discrepancy of age estimates and errors in spatial distribution of separate glacial episodes. Glaciations began in the mid-Carboniferous (Lopez-Gamundi, 1997), and the belt was evidently widest in the terminal Late Carboniferous–initial Early Permian.[2] As many researchers believe now, the Early Permian glaciers were most widespread in the Asselian–Sakmarian time (Visser, 1990, 1994, 1996, 1997; Visser *et al.*, 1997; Frakes *et al.*, 1992; Eyles and Young, 1994; Crowell, 1995; Dickins, 1996; Eyles *et al.*, 1998; and others). Glaciation of that time spread over high and middle latitudes of South America, Africa (including Arabia and Madagascar), India, Tibet, and Australia. Its influence is also recorded in the Malacca–Burma block. Antarctica that surrounded the South Pole was probably entirely covered by ice. The glacial belt was sometimes about 45°–60° wide (Figs. 1, 2). Ice shields and mountainous glaciers left numerous marks of glacial exaration (striated glacial bed with characteristic structures, trough valleys, fiords), different tillites, fluvioglacial, lacustrine–glacial, and glaciomarine sediments in these continental areas. Glaciomarine sediments are particularly widespread, because they were rapidly buried in vast

[2] The width of glacial belts and the total ice volume on the Earth are not strictly proportional. Judging from the eustatic and oxygen-isotope curves, as well as from mathematical modeling and calculated paleotemperatures (Gonzalez-Bonorino and Eyles, 1995), the total ice volume was maximal at the beginning of the Middle Carboniferous.

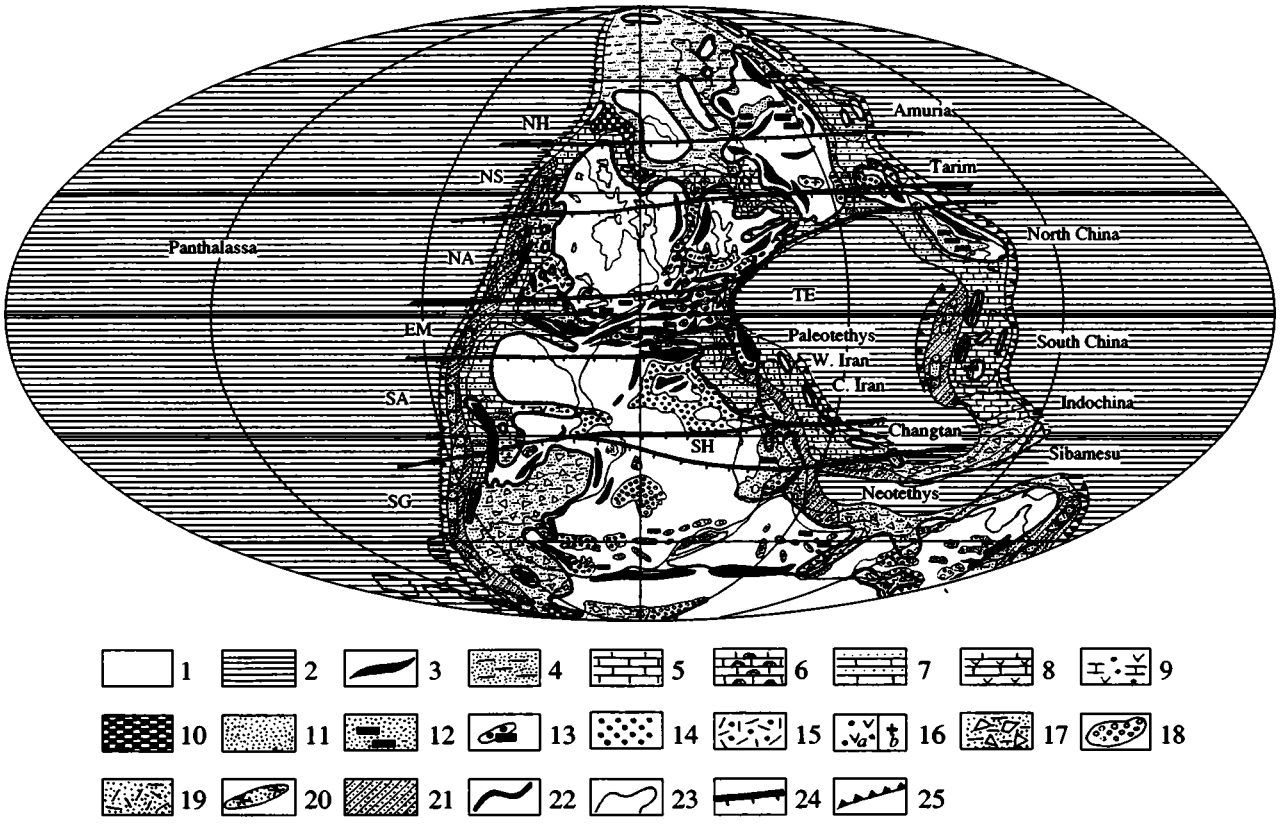

Fig. 1. Lithologic–paleogeographic map for the Asselian–early Sakmarian period of the Early Permian: (1) land; (2) oceans; (3) mountains; (4) shelf seas with terrigenous sedimentation; (5) carbonate platforms; (6) carbonate reefal deposits; (7) shelf seas with terrigenous–carbonate sedimentation; (8) evaporite–carbonate platforms; (9) inner sulfate–carbonate and sebkha basins that accumulated terrigenous red beds; (10) anoxic (black shale) basins; (11) epicontinental and coastal alluvial and alluvial–lacustrine basins in humid zones; (12) coal-bearing basins; (13) red bed terrigenous coal-bearing basins; (14) epicontinental and coastal alluvial, eolian, and lacustrine basins with red beds in arid zones; (15) basins of volcanogenic–terrigenous continental sedimentation; (16-a) epicontinental and coastal alluvial, lacustrine, sebkha, and saliferous basins with terrigenous red beds and gypsum-bearing strata; (16-b) saliferous basins; (17) glaciomarine and glacial deposits; (18) continental glacial deposits; (19) epicontinental basins with gray terrigenous and volcanogenic sediments; (20) terrigenous–volcanogenic complexes of island arcs; (21) turbidite troughs; (22) past coastal lines; (23) modern coastal lines, boundaries of sedimentary basins and lithologic–paleogeographic zones; (24) boundaries of sedimentation belts; (25) subduction zones. Letter symbols: (SG) southern glacial; (SH) southern humid, coal-bearing; (SA) southern arid; (EM) equatorial mountainous; (TE) tropical–equatorial; (NA) northern arid evaporite; (NS) northern semi-arid; (NH) northern humid, coal-bearing; (W. Iran) western Iran; (C. Iran) central Iran.

sedimentation basins in contrast to continental glacial deposits. Glaciomarine sediments accumulated under the influence of shelf glaciers, thawing waters, and icebergs. They were reworked to a different extent by the underwater colluvial processes. In the late Sakmarian–initial Artinskian, glaciers retreated everywhere and the glacial belt became narrower. Its northern boundary was located near the Antarctic Circle (Fig. 3). This interpretation is based on regional correlation and on faunal, palynological, and isotopic data obtained for the Dwyka Group of South Africa (Visser, 1990, 1997; Stollhofen *et al.*, 2000), several Upper Paleozoic glacial successions of Australia (Dickins, 1996; Eyles *et al.*, 1998; and others), and for the Itarare Group of South America (Franca, 1994; Santos *et al.*, 1996).

In the initial Early Permian, the Karroo, Kalahari, and Karasburg basins of South Africa accumulated a thick sequence of glacial deposits—the upper part of the Dwyka Group (up to 700 m thick).[3] Along periphery of the basins and on separating uplifts, the group is composed of continental glacial sediments, and mainly glaciomarine sediments are characteristic of central and southwestern parts of basins. Glaciers advanced in the basins probably twice (in the Asselian and Sakmarian) and subsequently retreated (in the terminal Asselian–initial Sakmarian and at the beginning of the Artinskian Age). The last signs of ice rafting are recorded in South Africa at the base of the Ecca Group and in the lower part of the Prince Albert Formation that is likely of the Artinskian age (Visser, 1994). The low degree of terrigenous material weathering in overlying deposits allows some authors to assume that some glaciation centers

[3] Hereinafter, we follow the stratigraphic scheme elaborated for South Africa by Visser (1990, 1997).

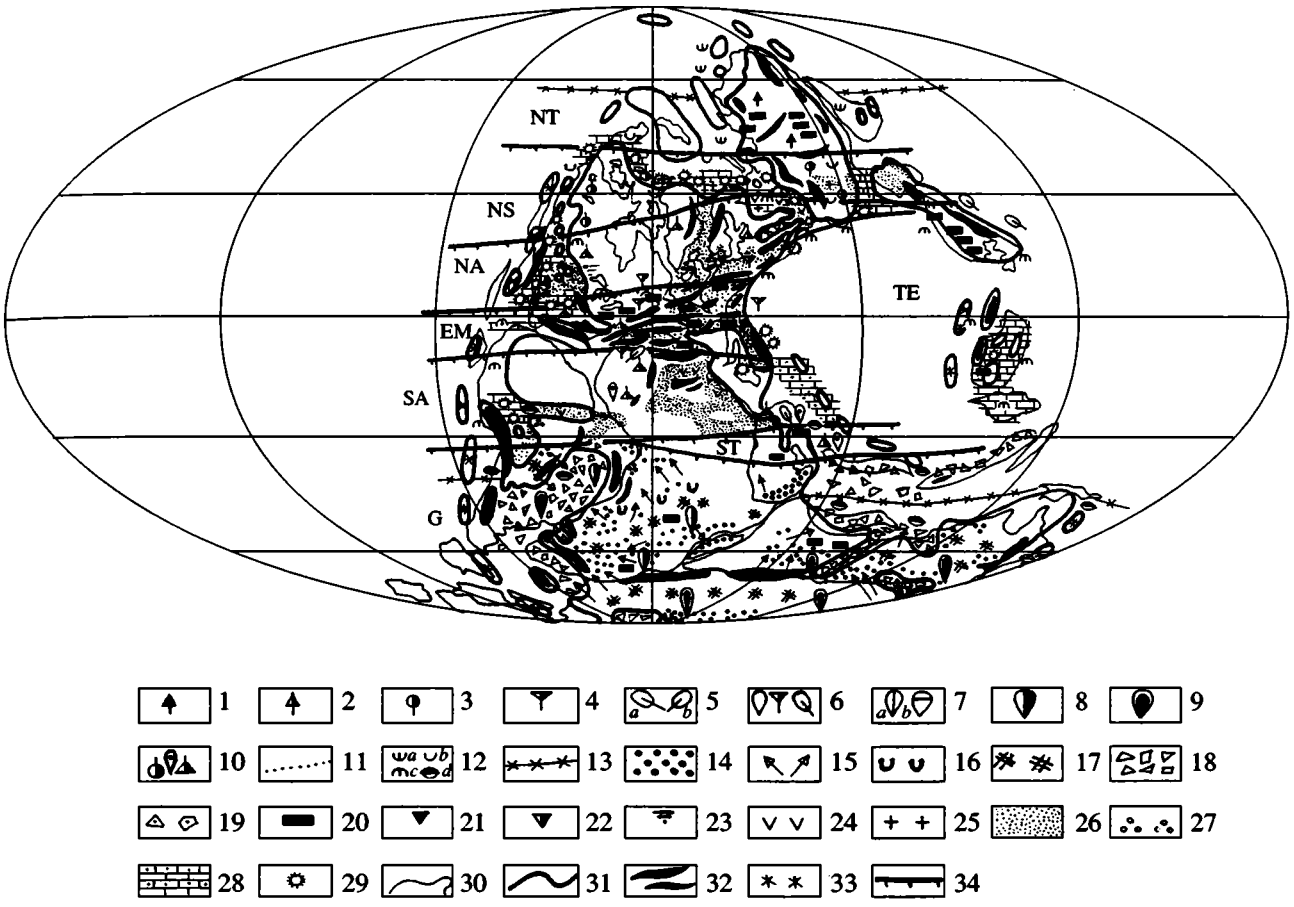

Fig. 2. Climatic zoning of the Asselian–early Sakmarian time with paleontological climatic indicators: (1) cold-resistant North Siberian vegetation, (2) moderately cold-resistant Siberian vegetation, (3) thermophilic Subangarida vegetation, (4) thermophilic Euramerian vegetation and also vegetation similar to the Euramerian one in the Arabian and Burma–Malacca regions, (5) thermo- and hygrophilic vegetation of the Cathaysian type in (a) North and (b) South Chinese floral regions, (6) thermophilic vegetation of the Southern Hemisphere (Gondwanan vegetation with elements of the Euramerian and/or Cathaysian floras in Africa), (7) moderately thermophilic vegetation of the Gondwanan Kingdom in (a) Australian–American and (b) Argentinean–Brazilian regions, (8) moderately cold-resistant vegetation of the Gondwanan Kingdom (*Glossopteris* floral region), (9) cold-resistant vegetation of the Gondwanan Kingdom (*Gangamopteris* floral region), (10) main localities of palynoflora in corresponding phytochores, (11) boundaries of some phytochores, (12) assemblages of shelf invertebrate fauna (after Grunt, 1995; Grunt and Shi, 1997) of (a) high-boreal, (b) low-boreal, (c) tropical (paleoequatorial), and (d) notal types, (13) northern and southern distribution boundaries of Early Permian conodonts (after Wardlaw, 1995); (14–29) lithological indicators: (14) mainly continental glacial and interglacial deposits, (15) directions of glacier movements, (16) fossil trough valleys, (17) glaciation centers, (18) mainly marine and glaciomarine deposits, (19) assumed glacial deposits, (20) coals; (21) bauxites, laterites, (22) humid soils, (23) carbonate, red, and variegated (partly gleyed) soils and calcretes, (24) gypsum and anhydrite, (25) stone, potassium, and other salts, (26) arid red beds, (27) eolian deposits, (28) carbonate platforms, (29) reefs (after Kiessling *et al.*, 1999); (30–33) some geographic elements: (30) modern shore lines, (31) past shore lines, (32) mountains, (33) volcanic chains; (34) boundaries of climatic belts. Letter symbols for climatic belts: (G) glacial, (SCT) southern cool-temperate, (NT, ST) northern and southern temperate, (NS, SS) northern and southern semiarid, (NA, SA) northern and southern arid, (EM) equatorial mountainous, (TE) tropical–equatorial.

could exist longer, up to the Kungurian Age (Visser and Young, 1990; Visser, 1994). The influence of different glaciation centers changed with time. In the Asselian Age, sedimentation was mainly affected by ice shields located in the northeast (Transvaal, Botswana, and Namibia), but later on, most influential were glaciation centers located in Antarctica, east and southeast of the South African Republic (Visser, 1997; Visser *et al.*, 1997). Simultaneously, glaciers advancing in the west–northwestern direction along trough valleys from the Namibian glaciation centers crossed ancient rises of western Namibia to reach the eastern flank of the Parana basin in South America. This is evident from preserved pre-glacial topography, glacial bed structures, and boulder orientations in tillites (Martin, 1981; Santos *et al.*, 1996). Marginal parts of the Parana basin accumulated tillites, fluvioglacial and lacustrine–glacial deposits; ice-rafted deposits are present in the central basin parts.

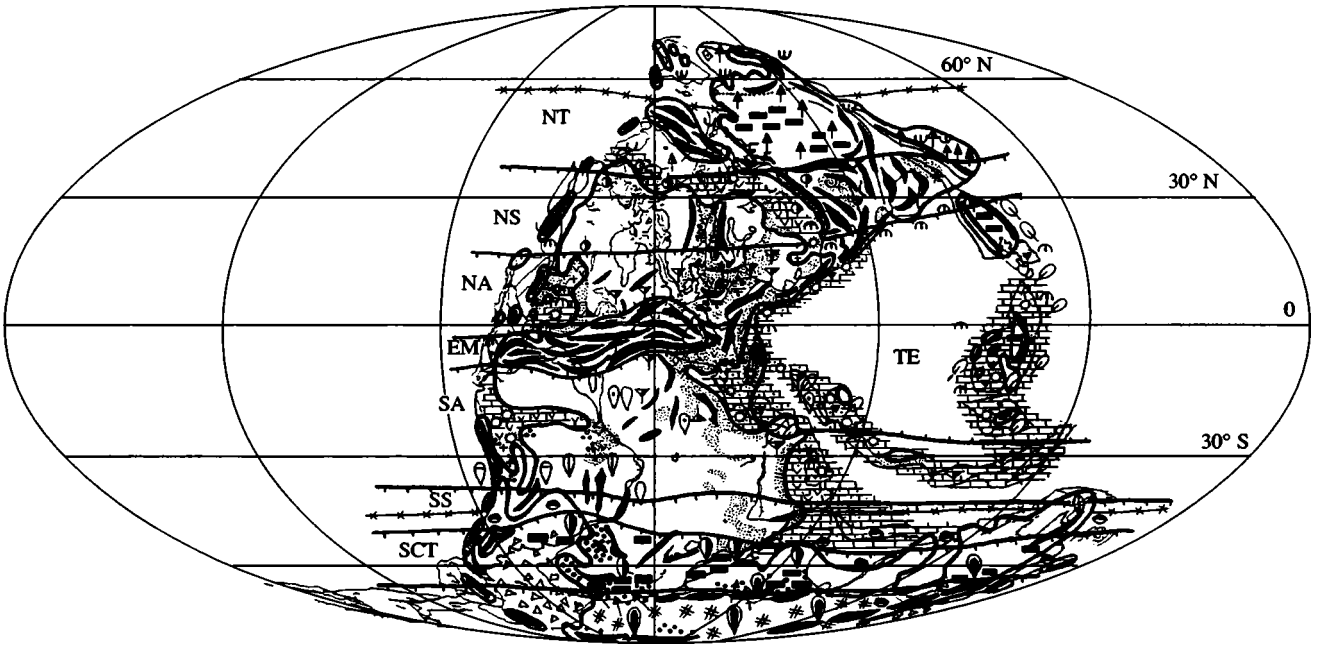

Fig. 3. Climatic zoning of the late Sakmarian–early Artinskian time (symbols as in Fig. 2).

During the Asselian–Sakmarian maximum, glaciers of Africa reached Gabon (Hambrey, 1981), Zair (Cahen and Lepersonne, 1981), and the southern Arabian Peninsula (Levell *et al.*, 1988; Alsharhan and Nairn, 1995; Stephenson, 1999). Thus, the northern boundary of the glacial belt in Africa was located near the paleolatitudes of 30°–40°S.

In Australia, the Lower Permian glacial deposits are preserved in numerous sedimentary basins distributed from the eastern coast of the continent and Tasmania in the south to the Jose Bonaparte Bay in the north (Lindsay, 1997). According to many researchers, Permian glaciations in Australia began in the Sakmarian and lasted, decreasing discretely during the Artinskian and Kungurian ages, at lest until the end of the epoch (Eyles *et al.*, 1998) and further, in the Kazanian (Crowell and Frakes, 1971; Frakes 1979), and even Tatarian time (Herbert, 1981; Veevers *et al.*, 1994). Other researchers suggest that glaciation in Australia was of the Asselian–early Sakmarian age, and that ice covers did not exist after this period (Dickins, 1996; Lindsay, 1997). In opinion of Dickins, the period from that glaciation to the Kazanian Age corresponded to fourfold alternation of cooling and warming phases. Cooling episodes are evident from dropstones (evidence of iceberg or seasonal ice rafting) and glendonites, while warm periods left behind kaolin, bauxites, and Tethyan invertebrates in deposits (Dickins, 1996). The contrasting character of these indicators and abundance of ice-rafted dropstones in Artinskian and Kungurian strata (Spry and Banks, 1962; Eyles *et al.*, 1998) cast doubt on the conclusion that glaciations terminated in Australia before the late Sakmarian time.

During the periods of maximum glaciation, ice shields and mountainous glaciers evidently covered the greatest part of Australia (Crowell and Frakes, 1971; Frakes, 1979) and probably advanced to the northern coast of the continent. Most significant glaciation centers were in the western, central, and southern parts of Australia (Lindsay, 1997) and in the Great Artesian Basin (Frakes, 1979). A thick ice sheet covered mountains along the eastern coast of the continent. Structures of glacial beds and tillites imply that glaciers advanced into the southern coastal area of Australia from the southeast, i.e., from Antarctica (Lindsay, 1997; Bourman and Alley, 1999). Antarctica supplied southeastern Australia mainly with the ice-rafted material. The glacial material was partly transported also from local mountainous centers (Eyles *et al.*, 1998). From northwestern and northern Australia, glaciers apparently advanced into sea basins of the Himalayas, Tibet, and, probably, the Malacca–Burma block, transporting the coarse ice-rafted material in these areas (Lindsay, 1997; Wopfer and Gasshyap, 1997).

In Australia, short climatic oscillations of different ranks complicated the main glacial periods, which lasted as long as several millions of years. For instance, four to five thick glacial members separated by interglacial deposits are registered in the Asselian Lyons Formation of glaciomarine sediments on West Australia (Condon, 1967; Dickins, 1985). Every glacial member consists of four to six smaller glacial submembers also separated by interglacial deposits (Condon, 1967). The

average duration of episodes corresponding to glacial and interglacial members and submembers did not exceed 1.00–1.25 and 0.18–0.46 m.y, respectively (Chumakov, 1985; Veevers and Powell, 1987). Some researchers report on signs of glaciation events in sections of southeastern Australia, which resemble the Pleistocene Heinrich events (Eyles *et al.*, 1997). Glacial oscillations are registered also in the Tillite Wynyard Formation of Tasmania that is composed of nine tillite members separated by sediments with marine fossils (Spry and Banks, 1962).

In South America, the Upper Permian glacial deposits are known from several basins southward of 10°S (in modern coordinates). The largest Parana Basin in southern Brazil is as large as 1200 × 1200 km. The Itarare Group corresponds to Upper Paleozoic glacial deposits of the basin, which are up to 1400 m thick. The largest upper part of the sequence (locally as thick as 1200 m) is referred to the Lower Permian. Viewpoints on its stratigraphic range are different. At present, some researchers believe that it spans the Asselian–Sakmarian (Santos *et al.*, 1996; Stephenson, 1999) or Asselian–Artinskian (Franca, 1994) interval, while others suggest that uppermost beds are of the Kungurian age. Like the Lower Permian glacial sections in other continents, the Itarare Group demonstrates the repeated (at least sevenfold; Rocha-Campos *et al.*, 1999) alternation of continental and glaciomarine facies deposited during glacial and interglacial episodes; the latter were frequently responsible for the coal accumulation. South America hosted several centers of Permian glaciation. As was noted, glaciers advanced from the southeast, i.e., from South Africa, into the eastern Parana Basin and from the Asuncion uplift into its western areas (Franca, 1994). The occurrence of Permian glacial deposits in the northeastern part of the Brazil Rondonia State implies that glaciation centers also existed north of the Parana Basin (Rocha-Campos, 1981). Into the Sergipe-Alagoas trough (the San Francisku River mouth area), the glacier penetrated from the Kongo River basin of equatorial Africa, probably moving via Gabon (Santos *et al.*, 1996; Visser, 1997). As is evident from distribution of Permian deposits in South America, the glacier reached the paleolatitude of 30°S.

Signs of Permian glaciations are also rather abundant in southern Asia. They are recorded in northeastern and northwestern India (Radjustan), northern Pakistan, in the Himalayas (from Pakistan in the west to Assam in the east, Hambrey *et al.*, 1981), in many sections of the Tibet from the Karakorum Ridge in the west to the Saluin River in the east and from Lhasa in the south to foothills of the Kuen Lun Mountains in the north (Metcalfe, 1994). In the Hindustan Peninsula, glacial deposits are preserved in numerous grabens at the base of Sakmarian–Artinskian Talchir Formation (Chandra, 1992). These sections predominantly consist of continental facies: tillites, boulder conglomerates, and other glacial sediments. Locally, the sections are crowned with glaciomarine sediments. In many places, the Talchir Formation overlies the glacier bed with peculiar indications of glacial exaration. The striation on the bed and orientation of elongated rock fragments point, with rare exceptions, to the general northeastern movement of glaciers during the Permian (Ahmad, 1981), when they advanced toward the Himalayas where mainly glaciomarine deposits occur. In the eastern Hindustan Peninsula, orientation of striation implies that glaciers moved from Antarctica adjoining the peninsula in the south. All these data and replacement of continental glacial deposits in the Himalayas and Tibet with glaciomarine ones indicate, contrary to the widespread opinion that only local ice domes existed in the peninsula, the northeastward movement of glaciers from Antarctica, via India, into an epicontinental sea located in the marginal part of Gondwana.

The Upper Paleozoic glacial deposits are known in the Transantarctic Mountains, along periphery of the Ronne Glacier, and in the western Queen Mod Land in the Atlantic segment of Antarctica (Hambrey *et al.*, 1981). In the Transantarctic Mountains, glacial deposits are 300 to 1000 m thick and fill in several sedimentary basins. They are represented here by continental and glaciomarine facies. According to palynological data, their stratigraphic range corresponds to the Carboniferous–Early Permian. The occurrence patterns of glacial sequences and directions of glaciers responsible for their accumulation (Isbell *et al.*, 1997) allow an assumption that some of the basins originated in response to glacial erosion (Isbell and Cueno, 1996). The contemporary ice cover is an obstacle for the detailed reconstruction of Late Paleozoic glaciations in Antarctica. Nevertheless, the fact that glaciers from this continent reached South Africa, India, and South Australia in the Early Permian time implies two conclusions: first, the Antarctic continent experienced strong glaciation at that time, and second, the glaciation lasted here until the Artinskian Age at least, because Antarctic glaciers still reached South Africa in the west (Visser, 1997) and India in the east (Ahmad, 1981; Chandra, 1992) at the beginning of this age. Antarctic icebergs continued to transport intermittently coarse detrital material to southeastern Australia until the Kungurian Age (Eyles *et al.*, 1998) or even until the end of the Late Permian epoch (Veevers *et al.*, 1994).

Despite the prevalence of glacial settings in the climatic belt under consideration, its marginal periglacial and even central (during interglacials) zones hosted vegetation and bogs, which accumulated peat subsequently transformed into coal seams. The interglacial floras and coals are known in southeastern Australia and South America. It is assumed that the peat accumulated after glacioeustatic transgressions in coastal bogs and river floodplains during the postglacial isostatic emergence of the region (Santos *et al.*, 1996; Bustin, 1997) or glacioeustatic sea-level falls (Michaelson and Henderson, 2000). The interglacial climate was sufficiently cool, as it is evident from the cold-resistant affinity of the primary vegetation and from occurrence

of solifluction, ice wedges, and impressions of ice crystals in coal-bearing rocks, which reveal admixture of fresh feldspar grains and enclose coarse ice-rafted material (Bustin, 1997; Martini, 1997; Eyles *et al.*, 1998; Michaelson and Henderson, 2000).

Vegetation of interglacial periods and that of periglacial margins of the belt was very impoverished. The initial Permian in Australia (Retallack, 1980) and South America (Guerra-Sommer *et al.*, 1999) was a time of wide development of herbaceous "tundra" and shrub pteridosperms (*Botrychiopsis*). Later on, this flora was supplemented by and then replaced with the dwarfish arboreal vegetation of the "taiga" affinity. The latter consisted mainly of probably deciduous *Gangamopteris* that could grow in permafrost areas (Retallack, 1980). This vegetation is known under the name "*Gangamopteris* floral assemblage" (*Gangamopteris* florula according to Retallack, 1980; *Gangamopteris* region after Wnuk, 1996). In Australia, Antarctica, Africa, and India, the impoverished flora (Cúneo, 1996) consisted of cold-resistant plants locally associated with tillites (Chandra, 1992) and permafrost signs (Retallack, 1980). In India, *Gangamopteris* remains occur in association with conchostracan and small insects (Chandra, 1992), and in Antarctica, interglacial deposits enclose remains and fucoids of freshwater crayfish. Judging from habitat conditions of modern crayfishes, some researchers believe that temperature in rivers and lakes of Antarctica could be as high as +10–+20°C during the Permian interglacial summers (Babcock *et al.*, 1998).

Sufficiently warm summer seasons of interglacial phases is also inferred from finds of fossilized stumps up to 1m in diameter in the Transantarctic Mountains. These remains probably belong to *Glossopteris* and partly occur in their life-time position. They have wide (up to 14 mm) summer–spring growth rings. Numerous cell layers in every ring point to favorable conditions for these trees during their vegetative period that lasted, as is roughly estimated, 48 days (Francis *et al.*, 1994). According to palynological assemblages, the enclosing rocks are of the late Early Permian age. The area with wood remains was situated at that time at the paleolatitudes of 80°–85°, where the dark polar period lasted at least 140–165 days. It can be assumed therefore that the interglacial summer season in high latitudes of the Southern Hemisphere was mostly unfavorable for arboreal vegetation because of low temperatures. Periglacial conditions remained also for some time after the glacier retreat. They were appropriate for development of the *Gangamopteris* assemblage, ice rafting, bog formation, and peat accumulation (Chandra, 1992; Isbell *et al.*, 1997; Smith *et al.*, 1998). During the glacier retreating and permafrost degradation, the *Gangamopteris* floral assemblage gave place to the more thermophilic *Glossopteris* flora, which completely substituted the *Gangamopteris* flora in Australia only by the end of the Late Permian. The period when the *Glossopteris* assemblage became prevalent can be considered as a time of eventual degradation of the Early Permian glacial belt in Gondwana. In South America, particularly, in Patagonia, vegetation was more diverse and thermophilic since the beginning of the Early Permian (Cúneo, 1996) despite the proximity of the glacial belt and penetration of South African glaciers into the Parana Basin located slightly westward (Santos *et al.*, 1996). This anomaly was probably related to the insular position of Patagonia, otherwise this flora can be of a younger age. Some researchers argue that the Patagonia block was separated from West Gondwana and situated during the Permian within lower latitudes than it is now believed (Cúneo, 1996). In Brazil, the Lower Permian flora resembled in composition the *Glossopteris* assemblage (Wnuk, 1996), although *Gangamopteris* forms prevailed locally in this region as well (Rocha-Campos and Santos, 1981).

Sea basins adjacent to glaciation regions and flood areas of interglacial glacioeustatic transgressions were populated by peculiar cold-resistant *Eurydesma* fauna, as it is frequently termed. This fauna was widespread in Australia, South Africa, India, Himalayas, Tibet, and southern Afghanistan (Termier *et al.*, 1973). It is also known in South America (Amos and Lopez Gamundi, 1981; Santos *et al.*, 1996). Some representatives of the *Eurydesma* fauna occur in the southern Pamirs (Grunt and Novikov, 1994). The *Eurydesma* fauna represented an impoverished variant of the Notal (Gondwanan, Australian) fauna (Runnegar, 1984). It was lacking fusulinids, reef-forming brachiopods, Rugosa, conodonts, and other thermophilic groups and was frequently associated with glaciomarine sediments and glendonites.[4] Bipolar forms represented significant components of this biota (Runnegar, 1984; Shi and Grunt, 2000).

The southern cold–temperate humid belt. In the terminal Sakmarian–initial Artinskian, when the Gondwanan glacial zone became significantly narrower and retreated toward the Antarctic Circle, a narrow (10°–15°) belt with prevalent temperate to cool climatic conditions formed immediately north of the glacial zone (Fig. 3, SCT). Numerous Lower Permian coal deposits of Australia, India, South Africa, and South America are situated in this belt. Relatively small local glaciation centers occurred sometimes within this belt, and tongues of Antarctic glaciers simultaneously reached the belt from the south. As was mentioned, the terminal Sakmarian–initial Artinskian glacial episodes are recorded in South Africa (Visser, 1997). The formation centers of local glaciers coincided with elevated areas. The glaciers moved mainly in the northwestern direction to the Parana Basin leaving behind mostly glaciomarine sequences. During glacial episodes, Antarctic glaciers reached eastern India (Ahmad, 1981; Chandra,

[4] The Notal biota of the Permian time was of a low generic diversity in general. For instance, the number of its brachiopod genera was six to eight times lower than in the tropical biota. Even in eastern Australia, the generic diversity of Permian brachiopods and bivalves decreased twice in the north–south direction with parallel decline of Rugosa abundance until the extinction.

1992). Small glaciation centers could probably appear in mountains of eastern Australia. The adjacent Sidney Basin received, in addition, a large amount of the ice-rafted material from Antarctica (Eyles *et al.*, 1997).

Nevertheless, climate in the belt under consideration was temperate to cool and humid despite some glacial episodes. This is evident from a wide distribution of the early *Glossopteris* floral assemblage reviling a notable percentage of *Gangamopteris* forms. Judging from the leaf accumulation, stem dimensions, wood and root structures, and abundant coal seams, this vegetation can be classed with the boggy deciduous forests with fern and horsetail underbrush (Retallack, 1980; Chandra, 1992). In the more arid areas, conifers represented subordinate components. The *Glossopteris* forests were populated by amphibians, reptilians, and large neuropterous insects (Chandra, 1992).

The proximity of the glacial belts determined to a significant extent the climate in adjacent sea basins. They accumulated locally ice-rafted sediments and were populated by the Notal fauna of the *Eurydesma* type. Seas of western Australia and Himalayan India near the northern boundary of the belt were populated by more thermophilic fauna (Indian province of the Gondwanan paleogeographic region according to Runnegar, 1984).

The southern temperate humid belt of middle latitudes. The available data indicate that climate of the Asselian–initial Sakmarian period was locally temperate and humid in the low middle latitudes (30°–40°S) of the Southern Hemisphere (Fig. 1, SH; Fig. 2, ST). This was characteristic of relatively limited areas, which did not form a continuous latitudinal belt. Indications of such a climate are known only from the central part of the Arabian block, where terrigenous sequences enclose thin coal seams (Alsharhan and Nairn, 1995) and yield the Cathaysian palynological assemblage that show admixture of Euramerian and Gondwanan taxa (Utting and Piasecki, 1995). Coals and composition of palynological spectra point to a sufficiently humid and warm climate of the territory in question located on the leeward monsoon-influenced margin of Pangea. Westward, this narrow band of the temperate humid climate is not traceable, because the gypsum-bearing red beds and eolian deposits are distributed close to the development area of glacial deposits in South America (Figs. 1, 2). Accordingly, we may assume that the temperate humid belt was missing here, and the glacial belt immediately contacted in the north with the arid one.

The northern temperate humid belt of middle and high latitudes. In the Early Permian, the entire northern part of Pangea, from the Arctic coast to paleolatitudes of 35°–40°N, was occupied by a wide belt of arboreal vegetation that was of Siberian phytogeographic type and gave rise to an intense coal accumulation (Fig. 1, NH; Figs. 2, 3, NT). Preponderant arboreal taxa of the belt were cordaites whose pycnoxilic wood had distinct growth rings indicative of the temperate climatic conditions. Some researchers compare that Siberian phytogeographic region with contemporary Boreal forests and conditionally apply term "Cordaitean taiga" to its vegetation (Durante, 1995). The Amuria continent of formerly amalgamated Primor'e, Transbaikal region, and Mongolia was evidently located within the same belt north of Siberia proper. This is evident from a wide development of the Siberian flora in that continent (Durante, 1995) and from remains of highly boreal fauna (Grunt, 1995). Remains of Siberian-type flora are also known from the Sakmarian deposits of the Canadian Arctic Archipelago (Wnuk, 1996). Thus, the northern temperate belt extended into the western part of Pangea.

The Siberian segment of Pangea extended far to the north in the Early Permian time (75°N according to Scotese and Langford, 1995; 80°N according to Ziegler *et al.*, 1997, 1998; about 68°N according to Parfenov *et al.*, 1999). If these data are correct, a presumable width of the northern temperate belt could be as wide as 35°–45°. According to Durante (1995), floral assemblages of a uniform Siberian composition occur up to the Okhotsk massif that was located near the northernmost boundary of Pangea. Nevertheless, some researchers tend to distinguish the South and less diverse North Siberian floras (Wnuk, 1992). Sea basins adjacent to the northern part of the belt lacked fusulinids (Leven, 1993) and conodonts (Wardlaw, 1995). Taking into consideration the aforementioned facts and a large width of the belt, we may assume that climate within the belt was variable and noticeably cooler in its northern areas. There are some indications of ice-rafted or glacial deposits in the Lower Permian sections of Siberia (Bobin, 1940; Andrianov, 1966), although their origin is debatable (Vikhert, 1957; and others) and unproven so far (Chumakov, 1994). Using these data, it is impossible do define the northern glacial belt, and only local glaciers, if there were any, could sporadically develop during the Early Permian in northern Siberia, where the temperate cool climate dominated.

Shelf seas of the belt under consideration were populated by the highly boreal invertebrate fauna (Grunt, 1995) that was of a low diversity and lacked colonial corals, fusulinids, and other thermophilic forms, though it shows presence of bipolar taxa (Ustritskii, 1993; Shi and Grunt, 2000; and others). Only areas near the southern boundary of the belt, in its western part, hosted more thermophilic low-boreal assemblages (Grunt, 1995).

The southern temperate semiarid belt of middle latitudes. In the terminal Sakmarian–initial Artinskian, a narrow semiarid belt 5°–15° wide appeared between the paleolatitudes of 40°–45° and 50°–55°S, where land areas were formerly ice-covered and begun to accumulate terrigenous, usually slightly calcareous red beds (Fig. 3, SS).[5] Significant amid them were alluvial

[5] Considering arid zones of the past, we call them "semiarid" following the definition and classification accepted in maps published by UNESCO (Shantz, 1958).

and lacustrine, sediments and fan deposits of seasonal flows. Sedimentation settings favorable for coal accumulation were missing. These features suggest that climate in the belt was arid in general, with short humid seasons responsible for appearance of branching and wandering short-term flows, and lakes. Proportions of alluvial and lacustrine facies in sections are variable, implying annual and secular moisture variations. The temperature was probably moderate. This can be inferred from the belt proximity to glaciation centers that episodically formed in the neighboring temperate cold belt and also from the occurrence of ice-rafted deposits in the Parana Basin (Santos *et al.*, 1996). This conclusion is also supported by terrestrial *Glossopteris* flora found in land areas (Wnuk, 1996), by wide development of Notal fauna in shelf seas (Grunt, 1995), where conodonts were scarce or completely absent (Wardlaw, 1995), and by the intermediate position of the belt between the temperate cold and typical warm arid ones (Fig. 3).

The northern warm semiarid belt of low latitudes. In distinction from its southern counterpart, the northern semiarid belt of middle and low paleolatitudes was sufficiently wide (15°–20°). During the Asselian Age, it was situated between 40°–45° and 15°–30°N (Figs. 1, 2, NS), while in the terminal Sakmarian–initial Artinskian it slightly widened southward, particularly in the Eurasian segment (Fig. 3). Later on, in the Kungurian Age, the belt shifted slightly northward. Land areas of the belt accumulated terrigenous, frequently calcareous alluvial–lacustrine sediments. Their sequences along coasts of lakes and epicontinental or shelf seas enclose sebkha deposits and caliche. In southern Kazakhstan, they host continental sulfate–sodium salts pointing to alternation of warm humid and cold arid seasons. As Zherebtsova (1977) estimated, the salts' formation temperatures were 20 to 30 and –5 to –15°C during warm and cold seasons, respectively. Such a wide temperature range suggests that climatic conditions within the belt could fluctuate from temperate cool to subtropical. More definite are some seasonal characteristics of the climate: the monsoon-type moistening and large seasonal variations of temperature typical of the continental climate.

Cool winter temperatures, as those indicated above, are now untypical of the same low latitudes and conflict with paleobotanical data characterizing the belt. In the Early Permian, vegetation of the belt was of the cordaites–conifers–pteridosperm type and included elements of tropical and Siberian floras usually associated with coal-free red beds. Proportions between tropical and Siberian forms in the flora were variable. Abundance of the latter increased in the Early Permian (Durante, 1995) that may be indicative of some cooling. In general, composition of this "Subangarida" flora implies the warm arid climate (Meyen, 1987; Durante, 1995) and open landscapes. This conclusion well agrees with the aforementioned properties of continental deposits and with composition of marine sediments in inner and marginal epicontinental seas of the belt. In the eastern Russian platform, southern Barents Sea, and northern Canadian Arctic Archipelago, sea basins hosted gypsum-bearing evaporite carbonate platforms, very spacious sometimes and fringed by reefal buildups. Sections in the southwestern Barents Sea (the Ottar basin and others) enclose salt-bearing deposits (Breivik *et al.*, 1995; Stemmerik and Worsley, 1995). The basins were mainly populated by moderately thermophilic fauna of the low-boreal type, and only the East European sea hosted the Mediterranean fauna (Grunt, 1995). The reef-building *Paleoaplisina* forms characteristic of northern areas of the belt (Beauchamp, 1995; Stemmerik and Worsley, 1995; Keissling *et al.*, 1999) could dwell, according to recent data, in moderately warm waters and were distributed mostly between the paleolatitudes of 25 and 45°N. Only in the western North American coast, they reached 15°N with the help, as is thought, of the cold anticyclonic Panthalassa coastal current (Keissling *et al.*, 1999). This assumption is well consistent with the southward advance of the Subangarida flora in this region (Utting and Piasecki, 1995). All these features are used to outline the northern semiarid belt in the North American segment of Pangea. In the south of the Eurasian belt segment, main reef-building organisms were represented by algae.

The northern arid belt of low latitudes. The southern part of the Laurasian segment of Pangea was occupied by a vast arid belt (Figs. 2, 3, NA).[6] During the Asselian Age, this belt extended from the paleolatitude 15°N almost up to equator in the west and from 25°–30° to the Paleotethys coast in the east. In the terminal Sakmarian–initial Artinskian, the northern arid belt of Pangea was located between equator and 15°–20°N. During this period, it widened southward and joined the southern arid belt along the eastern margin of the supercontinent (Fig. 3). In the terminal Permian, the belt again slightly widened northward. In the east, along the Paleotethys coast, the belt was narrower and probably pinched out (Figs. 2, 3). Land areas of the belt accumulated terrigenous, commonly gypsum-bearing sebkha and eolian (dunes) red beds. In the southwestern Moscow syneclise, Dnieper–Donets depression, and in western North America (Denver and Julesburg depressions, Supui Basin), saliferous sequences formed. In the Kungurian Age, areas and rates of salt accumulation in the northern arid belt significantly increased. Saliferous deposits accumulated in the Anadarko, Kansas, Julesburg, Williston, West Texas, and North Mexican, western North American, Central and East European basins (Zharkov, 1978). The initial thickness of the Kungurian saliferous sequence in the central part of the North Caspian depression (southern part of the East European basin) is estimated to be 4–5 km.

[6] We combined in this case the arid proper and extra-arid zones, because the available geological data do not permit their discrimination in line with classification used in maps published by UNESCO (Shantz, 1958).

The belt in question was characterized by the thermophilic xerophilous Euramerian vegetation that was relatively impoverished and largely composed of conifers (Meyen, 1987; Wnuk, 1995). The spores–pollen spectra from the belt commonly show presence of pollen of dessert and semidesert plants (Utting and Paiasecki, 1995). The climate aridity is also evident from sedimentation patterns in adjacent shelf seas, which hosted during the early Permian the accumulation settings of coastal gypsum-bearing sebkha deposits and spacious carbonate or evaporite–carbonate platforms (Zharkov and Chumakov, 2001). Remains of tropical fauna imply that sea basins were warm-water in general (Grunt, 1995).

The southern arid belt of low and middle latitudes. In the Southern Hemisphere, the arid belt represented a zone of development of terrigenous red beds, frequent gypsum-bearing sebkha sequences and eolian dunes. It was slightly wider that the arid belt of Northern Hemisphere and mainly occupied the low latitudes (between the paleolatitudes of 10° and 30°N; Figs. 1, 2, SA). As was already noted, it bordered the glacial belt in the west and the temperate one in the east. When glaciers retreated in the terminal Sakmarian–initial Artinskian period, the southern arid belt considerably widened toward the middle latitudes to be located between the paleolatitudes of 0°–15° and 40°–45°S. In South America, it included the vast saliferous Peruvian–Bolivian and evaporite Amazonian basins. The eastern part of the belt was occupied by the spacious Arabian and North Italian zones of evaporite–terrigenous–carbonate sedimentation (Zharkov and Chumakov, 2001). Thus, the southern arid belt became almost twice wider than its northern counterpart and was located farther from the equator (Fig. 3).

The plant and relevant palynological assemblages of the initial Permian, which are known from the belt, are of a mixed composition: the Euroamerian–Cathaysian in the north and Euramerian–Gondwanan in the south (Broutin *et al.*, 1990; Wnuk, 1996). After the belt expansion, its phytobiogeographic features became rather intricate that was quite natural for the belt of huge dimensions. The northern part of the belt (North Africa) was populated by the Euramerian flora that mixed southerly (Central Africa) with the Gondwanan flora and slightly easterly (Turkey) with the Gondwanan and Cathaysian floras (Meyen, 1987; Broutin *et al.*, 1990; Wnuk, 1996). The last type of floras probably indicates some humidity growth in the eastern peri-oceanic part of Central Pangea. The southern part of the belt under consideration belonged almost entirely to the Gondwanan floral kingdom. This segment of the belt corresponds to the extended area with dispersed remains of the so-called Australian–African–American flora (Wnuk, 1996) that included arboreal ferns and lycopods, in addition to *Glossopteris* forms, and presumably was of the thermophilic type. Localities of this flora are known near areas of gypsum-bearing sebkha and dune sand accumulation sites. This suggests that its development was connected with oases. The so-called "Argentinean–Brazilian desert flora" reflects cooler conditions. (Wnuk, 1996). It grew in the extreme southwest of the southern arid belt near the saliferous and gypsum-bearing basins along the boundary with the temperate semiarid belt. Some coniferous remains were found in this flora from eolian sands. Thus, the southern arid belt of Pangea shows a certain climatic differentiation from the warm climate in the north to the temperate warm or even temperate one in the south.

The eastern continuation of the southern arid belt is traceable in the southern part of the Tethys, where microcontinents of the southeastern part of the Kimmerian arc host red beds and remains of the Euroamerian and Australian–African–American floras (the Burma–Malacca and New Guinea blocks, respectively; Wnuk, 1996). Shelf and inner seas adjacent to the belt were characterized by development of coastal gypsum-bearing sebkhas (Zharkov and Chumakov, 2001). Biogeography of marine basins was as intricate as in continental areas. The tropical (paleoequatorial) fauna was characteristic of the Andean zone along the western margin of the belt, whereas the Notal (Gondwanan according to Grunt, 1995, and Grunt and Shi, 1997) vegetation was distributed easterly, in the Amazonian basin and near the Arabian block.

The equatorial belt of mountainous climate. The huge belt of Hercynian mountains formed as a result of collision between Laurasia and Gondwana and crossing Pangea almost from ocean to ocean was located in the central part of supercontinent between the northern and southern arid belts. In the Early Permian, it mainly occupied southern, near-equatorial paleolatitudes (5–10°S). Narrow sea gulfs still existed in the belt at the beginning of the Permian (Autunian or early Wolfcampian periods). The Val-Verde sea trough that accumulated terrigenous deep-water sediments, partly turbidites (Oriel *et al.*, 1967), stretched along the northern front of the Wichita-Marathon orogenic belt. Northward, the Eastern shelf corresponded to a shallow sea basin, which accumulated clayey and carbonate sediments, whereas coal-bearing and fluvial terrigenous deposits were characteristic of its eastern coastal margins (Oriel *et al.*, 1967). The area south of the Wichita-Marathon orogenic belt was occupied by a fore-arc sea that stretched probably to southern Arkansas and northern Louisiana separating the continental blocks of South and North America (Wickman *et al.*, 1976; Frazier and Schwimmer, 1987). In the eastern margin of the mountainous belt, the sea entered North Africa spreading over the northern part of the Mauab-Ramzes Basin (southern Tunisia), where thick sandy–clayey and carbonate sediments are deposited (Hoffmann-Rothe, 1966; Vysotskii *et al.*, 1973). During the Asselian Age, the area northwest of the belt was occupied by a spacious alluvial–lacustrine boggy plain that was populated by freshwater ostracodes, fishes, and amphibians, and where peat subsequently metamor-

phosed to coal accumulated. The occurrence of brachiopods suggests episodic sea invasions or a partial salinization of waters in the plain. In dry seasons, the plain accumulated, particularly in its southern part, alluvial red beds (Berryhill, 1967).

Piedmonts and numerous intermontane depressions in central and eastern parts of the mountainous belt represented deposition sites of volcanogenic, alluvial–lacustrine, gray-colored deposits and red beds. In the Autun depression of France, the Asselian (Autunian) deposits are represented by gray to black calcareous and bituminous shales intercalated with fluvial, partly variegated beds and volcanic ash layers, which yield the characteristic flora in association with freshwater algae whose accumulations are transformed into sapropel coals. These deposits accumulated under conditions of the hot and humid climate (Cassins *et al.*, 1995; Châteauneuf and Farjanel, 1989). Similar humid environments prevailed in the Pyrenees and southeastern France, where coal-bearing sequences are present. Intermontane depressions in the Iberian plate accumulated alluvial–lacustrine red beds alternating with gray to black terrigenous sediments and volcanics. Almost all depressions here bear coals (Cassins *et al.*, 1995). In the Southern Alps of Italy, the Upper Permian deposits fill in several deep grabens (Orobic, Val-Trompia, Tione, Bolzano, and others basins). They are represented by volcanogenic, fluvial, and lacustrine rock complexes accumulated under conditions of the warm and temperate semiarid climate with alternating warm and dry periods (Cassins *et al.*, 1995). Similar sedimentation environments were also characteristic of deposition sites of continental coal-bearing red beds in southern areas of the mountainous belt, namely in the Tiddas, Argana, Ourica, Chougrane, Haouz, Oued-Zfe, and other intermontane basins of northwestern African (Conrel *et al.*, 1991; Wartiti *et al.*, 19990).

Thus, it can be stated that the local climate in piedmonts and intermontane depressions was very variable at the beginning of the Early Permian and depended on orographic peculiarities of the belt. Coal-bearing deposits frequent in depressions point to alternation of predominant warm and humid climatic phases with semiarid periods. The Autunian red beds could accumulate in areas protected from rains by mountains. Alternation of semiarid and humid climate likely was typical of the eastern and western margins of the belt. The vertical climatic zoning of mountainous forests, meadows, and steppes, like that in modern equatorial and tropical belts, was characteristic of mountains. The Euramerian flora from Autunian deposits includes the Gondwanan and abundant thermo- and hygrophilic equatorial Cathaysian elements (Broutin *et al.*, 1990; Wnuk, 1996). Later on, during the late Sakmarian–Artinskian time, remaining sea basins closed in response to the general uplift of the orogenic system, and the mountainous belt of Central Pangea that probably resembled the Himalayas terminated its formation (Scotese and Langford, 1995). Simultaneously, climate changed to semiarid and arid. Piedmonts and intermontane depressions mostly accumulated red beds frequently enclosing horizons of carbonate, partly gleyed soil, caliche, and gypsum (Mader, 1992; Cassins *et al.*, 1995). It can be assumed that climatic zoning in the region approached that of the intracontinental arid type (Zharkov and Chumakov, 2001).

The tropical–equatorial hot humid belt. This belt was spread over the Paleotethys, its coasts, Cathaysian microcontinents, and the northern part of the Cimmerian arc that bordered the ocean in the east and south (Figs. 1, 2, 3, TE). In its widest eastern part, the belt extended almost from 30°N to 30°S occupying all tropical and subtropical paleolatitudes. From the formal standpoint, this isometric region can be termed the belt conditionally, but in essence, its vast territory with sufficiently uniform climatic, sedimentation, and biogeographic environments did not differ from other climatic belts, because its main characteristics were also determined by the latitudinal geographic position and orography. The belt included the eastern insular and coastal humid zones of the tropical (northern and southern) and equatorial belts, which were closer to each other than to western continental segments of all three belts.

Abundant signs of hot and humid climate are observable on land areas of the former tropical–equatorial belt, e.g., coals, bauxites (Zharkov and Chumakov, 2001), and diverse remains of Cathaysian vegetation (Wnuk, 1996). The latter included typical plants of humid tropics: arboreal *Lepidodendron* and Equisetales, and also *Gigantopteris* representing probably lianas or climbers (Ziegler, 1990). The Cathaysian phytochore was not uniform, and this is quite natural in view of its tremendous width and predominant archipelago-type land patterns. Three floral regions of different latitudinal positions are recognized now within the Cathaysian arc: the North and South Chinese regions corresponding to opposite margins of the arc and the intermediate narrow Ksu-Huan-Yu region (Wnuk, 1996). This phytogeographic differentiation likely developed during the Early Permian (Wang *et al.*, 1998) has reflected both the isolated insular character of floras and their different latitudinal positions within the belt. This is suggested by the fact that the low-latitude South Chinese paleofloral region (South China and Indochina microcontinents) stretched in the latitudinal direction far westward up to Central Iran and northeastern Arabia, which were remote from the Cathaysian arc farther than the Cathaysian blocks from each other. This assumption is supported by some paleobotanical data implying the increasing humidity of the Permian climate within the Cathaysian arc in the north–south direction (Zo, 1999). The western part of the belt under consideration, where different phytochores joined, hosted the mixed Euroamerian, Gondwanan, and South Cathaysian floras (Broutin *et al.*, 1990; Wnuk, 1996).

Spacious carbonate platforms with reefal buildups and diverse Tethyan fauna were characteristic of seas

adjacent to the belt in question (Leven, 1993; Grunt, 1995; Grunt and Shi, 1997). Noteworthy were carbonate platforms and reefs widespread during the Asselian in the low and, partly, middle paleolatitudes of the Northern Hemisphere (up to 45° and 40°N), but missing south of 27°S (Figs. 1, 2) and 15°S (Kiessling *et al.*, 1999) in the Southern Hemisphere. Thus, the latitudinal distribution range of carbonate platforms and reefs in the latter was 2–2.5 times narrower than in the Northern Hemisphere during the Asselian glacial maximum and than in the Southern Hemisphere during the later period of the Early Permian (Figs. 2, 3). This indicates that influence of the glacial belt was noticeable up to the southern part of the tropical–equatorial belt. It could be responsible for some general cooling in the region and/or for penetration of cold and slightly turbid waters here.

PECULIARITIES OF THE EARLY PERMIAN CLIMATE

As one can see from previous sections, the Early Permian climate was highly variable and dynamic. We will characterize only its main peculiarities and changes that occurred throughout this epoch.

Global warming. The retreat of glaciers from middle latitudes toward higher ones in the late Sakmarian–early Artinskian time, when the warm arid belt and evaporite–carbonate platforms spread over former glaciation areas, indicate a rapid planetary warming. The reduction of glaciated areas and development of glacioeustatic transgressions decreased the planetary albedo and resulted in warming that accelerated up to the moment when glaciers entirely retreated to high latitudes. After this event, warming decelerated. Small glaciation centers, which sometimes grew in size due to different-rank climatic oscillations, were preserved in southern high latitudes up to the terminal Early Permian. The probable causes responsible for the warming will be discussed in out next paper. Now, we want just to note that, judging from the discussed materials, changes in the Pangea latitudinal position and in proportions between land and sea areas could hardly represent the main factors, because they were insignificant. According to paleomagnetic data (Scotese and Langford, 1995; Ziegler *et al.*, 1997, 1998), the northward displacement of supercontinent during the Early Permian corresponded to several degrees only, as is estimated. The subsequent rise of Pangea and reduction of epicontinental seas should cause the global cooling. The late Artinskian–Kungurian time was probably the warmest period in the entire Early Permian. Nevertheless, abundant indications of the iceberg and ice rafting in southeastern Australia suggest that cool climate in southern polar latitudes lasted up to the end of this epoch.

Global aridity growth. Global climatic changes of the Early Permian time have not been limited by warming. Since its formation time, semiarid and arid climatic environments were characteristic of low latitudes of Pangea. Factors responsible for this were the huge dimensions of supercontinent and emergence of large Hercynian mountainous belts and ranges, many of which extended along continental margins, particularly in the Laurasian segment. Both factors hindered the moisture transfer from oceans into internal areas of the supercontinent. The aridity of Pangea strengthened with time during the Early Permian, as it is reflected in extension and migration of arid and semiarid belts into the middle latitudes. The aridity grew in the equatorial mountainous part of central Pangea as well. Undoubtedly, one of factors responsible for this was global warming that increased evaporation, particularly in low and middle latitudes, where moisture was insufficient. The period of maximal extension of arid belts coincided probably with accumulation time of Kungurian salts and sulfates. The successive regression of inner seas of Pangea and relevant reduction of moisture sources could represent another cause of the arid climate advance. It can also be assumed that the retreat of large ice shields into high southern latitudes in the late Sakmarian–early Artinskian time resulted in reduction and retreat of corresponding annual stable high-pressure zones. This was evidently accompanied by enlargement of adjacent low-latitude Hadley cell and by migration toward poles of high-pressure areas characteristic of tropics and determining development of arid and semiarid belts.

The monsoon role. The existence of spacious Pangea land blocks predetermined an importance of seasonal areas with high and low air pressure and relevant influence of monsoons on the climate formation in both hemispheres (Parrish, 1995). This influence should be most significant in the Southern Hemisphere, where the compact block of the Gondwana land was almost 1.5 times larger in size and across (along 30°) than the present-day Eurasia. The Laurasian segment of Pangea was lesser only slightly. The humid monsoon climate was most distinct in areas adjacent to the Paleotethys ocean. The proximity of the warm ocean located in tropical and equatorial latitudes to huge blocks of the leeward land should stimulate development of monsoons and enhanced their influence. Taking into consideration this fact and mathematical models, some researchers termed these atmospheric events as megamonsoons (Kutzbach and Gallimore, 1989). However, on the way to Laurasia, these monsoons met mountainous belts and marginal ranges fringing the continent in the south, southeast, and northeast. As a result, monsoons connected with the lower troposphere lost the bulk of transported moisture on the leeward slopes of mountains. Therefore, climate in areas protected from rains and in tropical latitudes of Laurasia was arid (Figs. 2, 3). Indications of the intermittent summer moistening observable in deposits of the northerly semiarid belt allow however a suggestion that highly weakened monsoons reached episodically the internal areas of Laurasia. Signs of humid seasons are locally

registered also along the southern margin of Pangea. Nevertheless, the year-round high-pressure maximum over the glacial belt probably prevented, especially during the Asselian–early Sakmarian period, from a deep penetration of monsoons into its internal areas. Later on, monsoons could be kept away by tropical and high-latitude high-pressure zones, which existed in northern Gondwana and southern polar area, respectively.

Dynamics of climatic changes. An important feature of Early Permian climatic changes was their high dynamics. Two factors could be responsible for this: the general rapid warming and frequent different-rank oscillations that complicated the warming trend. The rate of global warming was highest during the Sakmarian Age, when glaciation in middle south latitudes terminated. A significant part of the territory that became free of ice was immediately occupied by the arid and semiarid belts (central South America, southern Arabia, Figs. 1, 2, 3). Further reduction of the glacial belt was more gradual. At the end of the Artinskian Age, this belt probably became extinct as a single whole, although glaciation centers that produced abundant icebergs repeatedly renewed in the Southern Polar zone during the Kungurian Age and even later (Eyles *et al.*, 1998).

Oscillations enhanced these dynamic climatic changes, particularly during glaciations. High expansion and even higher degradation (termination) rates of ice covers are well exemplified by Pleistocene glaciations. The available data show that similar short-period oscillations were characteristic of Early Permian glaciations as well (Chumakov, 1985; Dickins, 1985; Eyles *et al.*, 1998). As was noted, glacial oscillations recorded in most complete sections of Permian glacial deposits were of three ranks: with periods of hundred thousands, several millions, and several tens of million years (Chumakov, 1985; Veevers and Powell, 1987). High-amplitude climatic oscillations are also inferable from coal seams occurring sometimes amid glacial deposits (South America, southeastern Australia). They suggest that the glacial climate gave place to the temperate cool one and vice versa. Amplitudes of climatic oscillations within the ice-free areas were naturally less contrast than in glacial belts. They were probably manifested as moisture variations (Miller *et al.*, 1996; and others).

Asymmetry in the climatic zoning. Throughout the entire Early Permian, the Southern Hemisphere was colder that the Northern one. This is evident from the fact that all significant glaciations occurred only in the Southern Hemisphere. In this connection, succession of climatic belts was different in both hemispheres, and corresponding belts were of different width, remote from the equator for a different distance, and showed difference in many other parameters (radiation, seasonal barometric patterns, atmospheric precipitation, wind activity, and others). The asymmetry was particularly well manifested during the Asselian–Sakmarian glaciation (Figs. 1, 2). By the end of the Early Permian, it noticeably decreased (Fig. 3). Main causes responsible for the climatic asymmetry were different dimensions of land blocks and their asymmetric latitudinal position relative to the equator and poles (Chumakov, 1994). In the Early Permian, Gondwana located between paleolatitudes of 5° and 90°S was 1.6 times larger than Laurasia, and a half of its territory was located in high and middle latitudes. Laurasia was situated between the equator and 75°N at that time, and 67% of its territory corresponded to low latitudes. When glaciation zone decreased, distribution of land blocks remained almost unchanged, and climatic asymmetry substantially lessened with time. This suggests that asymmetric position of land masses results in strong climatic asymmetry only during glaciations.

Rearrangement of climatic zoning. The Early Permian climatic changes substantially rearranged the climatic zoning. Transformations were strongest in the Southern Hemisphere, where the glacial belt considerably (by 2.5 times and more) narrowed in the late Sakmarian–early Artinskian time and disappeared at the end of the Early Permian. In the Asselian–early Sakmarian time, the glacial belt bordered immediately the tropical arid belt in western Gondwana and the significantly reduced temperate belt in the east (Fig. 2). Reduction of the glacial belt was accompanied by formation of the semiarid and temperate cool belts between the arid and glacial belts. The temperate cool belt gradually expanding southward replaced the glacial belt at the end of the Early Permian. Simultaneously, semiarid and arid belts of both hemispheres became wider, and aridity increased as well in the equatorial mountainous humid belt. As a consequence, the planetary climatic zoning was drastically transformed.

Evolution of global climate. The climatic zoning is a most distinct reflection of the global climate, and we can confidently speak therefore about the substantial changes in the global climate during the Early Permian. In general, this period can be classed with the glacial epoch, because glaciers developed, although episodically, to the end of the epoch at least in the Southern Hemisphere. By the extent and character of glaciation, three stages in evolution of the Early Permian glacial climate can be distinguished. The Asselian–early Sakmarian climate was similar, in the extent of glaciation, to the Pleistocene one. In these cases, ice shields spread over high and middle latitudes of both hemispheres. Accordingly, it was the *climate of the glacial maximum*, as we can term it. In the popular literature, such glaciations are frequently called "great glaciations." Climate of the terminal Sakmarian–initial Artinskian time can be termed as the *climate of polar cups*, because glacial sheets were mainly confined to high latitudes. Climate of this type exists now and has existed in the Oligocene and in the second half of the Eocene, when first ice covers appeared in Antarctica. At the end of the early Permian, the cold temperate climate became prevalent in polar areas of both hemispheres. The southern polar cap degraded with preservation of small glaciation centers (most likely, episodic ones). This time can be consid-

ered as a terminal stage of the Gondwanan Glacioera and corresponding global climate as the *climate of cold polar areas.*

ACKNOWLEDGMENTS

We express our sincere gratitude to V.G. Ganelin, A.V. Goman'kov, M.V. Durante, P.K. Link, S.V. Naugil'nykh, and, particularly, E.Ja. Leven for their comments and valuable recommendations when we worked on the manuscript. The work was supported by the Russian Foundation for Basic Research, project nos. 99-05-64073 and 99-05-65004.

Reviewers A.S. Alekseev and E.Ja. Leven

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Stratigraphy and Geological Correlation, Vol. 10, No. 6, 2002, pp. 603–626. Translated from Stratigrafiya. Geologicheskaya Korrelyatsiya, Vol. 10, No. 6, 2002, pp. 82–105.
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New Stratigraphic Scheme of the Lower Aptian in the Volga River Middle Courses

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Received June 14, 2000; in final form, October 25, 2000

Abstract—Aptian deposits in middle courses of the Volga River have been under investigation for more than a century. Although stratigraphy was elaborated in detail for that region, new data allowed revision and further development of available stratigraphic schemes. Previously, the age of lower Aptian deposits was traditionally established based on changing ammonite assemblages of the family Deshayesitidae. The studied diverse assemblage of heteromorphic Ancyloceratidae, the habitants of relatively deep basin parts, made it possible to propose a new scheme of ammonoid zonation in the lower Aptian epipelagic deposits of the Russian plate. Many of identified ancyloceratids were established here for the first time. The analysis of coexisting deshayesitids and heteromorphs enabled a correlation of stratigraphic schemes for the monomorphic Deshayesitidae and heteromorphic Ancyloceratidae. The described generic taxa and species are *Volgoceratoides* I. Michailova et Baraboshkin, gen. nov., *V. schilovkensis* I. Michailova et Baraboshkin, sp. nov., *Koeneniceras* I. Michailova et Baraboshkin, gen. nov., *K. tenuiplicatum* (von Koenen, 1902), *K. rareplicatum* I. Michailova et Baraboshkin, sp. nov.

Key words: lower Aptian, Russian plate, Volga River middle courses, Cretaceous, stratigraphy, ammonites.

In the Russian plate, the most complete sections of the Aptian Stage are exposed within the Ul'yanovsk–Saratov syneclise in the Volga River middle courses. The lower Aptian part of the interval is best characterized by fauna and hence studied best. Nevertheless, there are some debatable points concerning both the biostratigraphy of lower Aptian deposits and the structure of the section itself.

A unique section of Aptian deposits was studied in 1995–2000 in the vicinity of the city park Yunost' of Ul'yanovsk. The section appeared due to stripping on the Volga bank for the construction of a new bridge. Other sections of the lower Aptian and Barremian–Aptian boundary interval were studies in landslides on the Volga right bank near the Village of Kremenki, downstream of Novoul'yanovsk, and near Sengilei (Fig. 1). Aptian sections located near Khvalynsk, Vol'sk, and Saratov were studied in 1998–2000. Stratigraphic description of the sections was carried out by Baraboshkin (Moscow State University, MGU) and his colleagues from the Research Institute of Geology of Saratov State University (RIG SSU) and from the Geological Institute of the Russian Academy of Sciences (GIN RAS). Ammonoids and their distribution through the section, as well as distinguishable zonal assemblages, were studied by present authors.

The new biostratigraphic scheme elaborated for the lower Aptian in the Volga River middle courses is based on a layer-by-layer study of sections, new findings of ammonites, and on revision of available collections donated to us by geologists from other institutions. It was this problem that the present work is devoted to.

INVESTIGATION HISTORY

The study of lower Aptian deposits is associated with many researchers who visited the Volga region between Ul'yanovsk and Saratov over 150 years. Sintsov, Sazonova, and Glazunova undoubtedly made a great contribution to the knowledge of this substage. Having started his works in the Saratov and Simbirsk regions in 1866, Sintsov published a series of papers on the geological structure of the regions in the 1870s–1880s. In earlier papers (Sintsov, 1870, 1872a), he described two new species *Ammonites trautscholdi* and *Crioceras tuberculatum*. Moreover, he focused his attention on *Ammonites deshayesi* (Leym.) found near Simbirsk and pictured in the work by Trautschold (1865).

Papers by Sintsov were preceded by Yazykov's works, the *Brief Review of Cretaceous Deposits of the Simbirsk Province* (1832) included. As Sintsov (1872b) reported, Yazykov kept fossils collected by him at the Museum of the Mining Institute, and in 1845, for the first time for the Simbirsk province, he pointed to the presence of ammonites *Ancyloceras* and *Hamites* in the sections he studied. Thus, it was Yazykov who pointed first to presence of Aptian ammonites in the region.

Preliminary brief reports on geological investigations were published by Sintsov in 1873, 1875, 1887

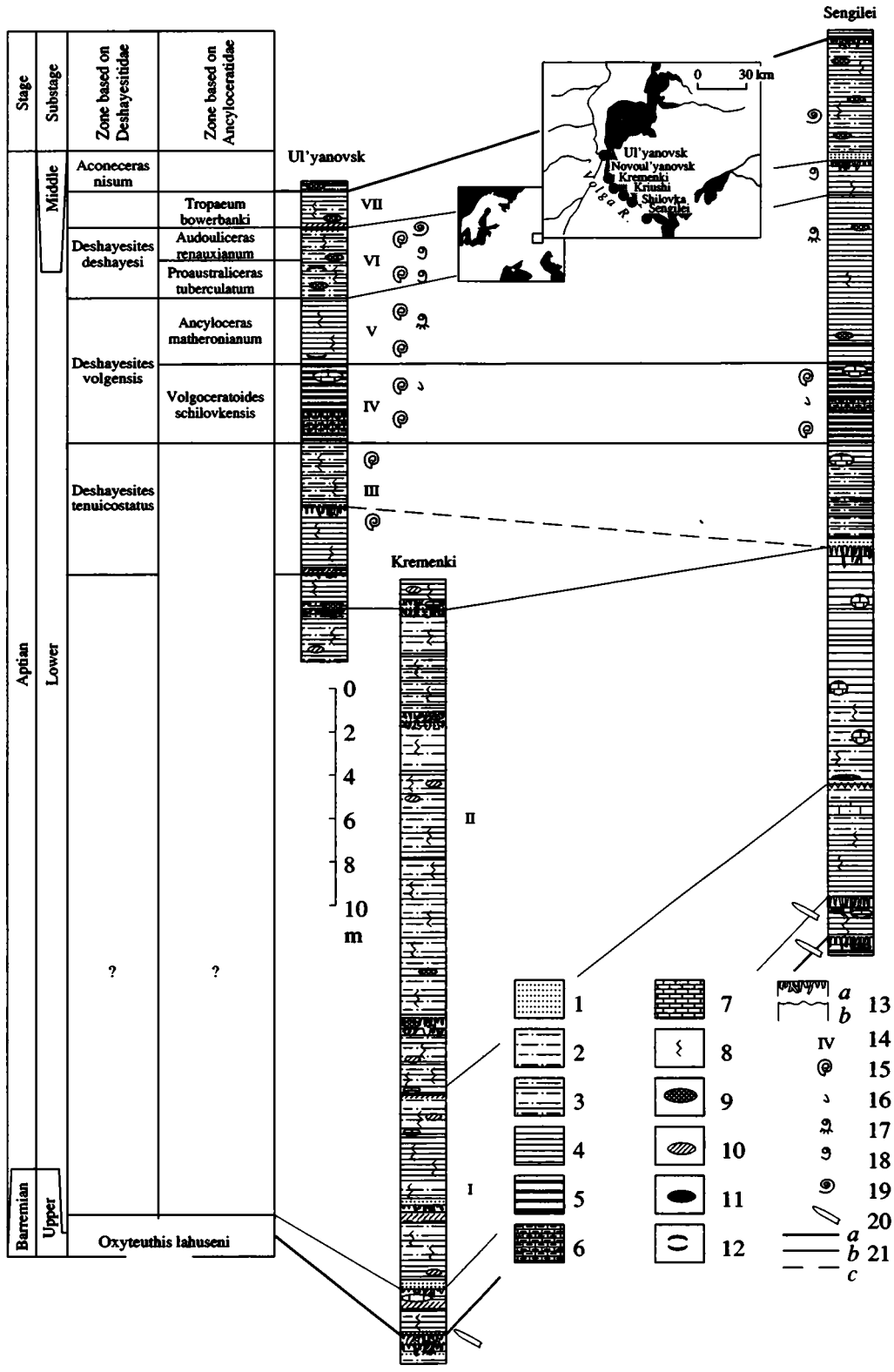


Fig. 1. Localities with ammonite fauna in the Ul'yanovsk region (inset map) and lithostratigraphy of the Ul'yanovsk (exposures near the new bridge), Kremenki, and Sengilei sections: (1) sand; (2) silt; (3) alternating sand and clay; (4) clay; (5) combustible shale; (6) clayey limestone; (7) limestone; (8) bioturbation; (9) siderite concretions; (10) sulfide concretions; (11) phosphorite; (12) shell detritus; (13) softground (*a*) and erosional (*b*) contacts; (14) member nos. after Baraboshkin, (1998); (15) *Deshayesites*, (16) *Volgoceratoides/Koeneniceras*, (17) *Ancyloceras/Lithancylus*, (18) *Audouliceras/Proaustraliceras*, (19) *Tropaeum*, and (20) *Oxyteuthis* faunas; (21) boundaries of stages and substages (*a*), members (*b*), and beds (*c*).

Table 1. Ammonite zonation in the lower Aptian Substage after Sazonova (1958) and Glazunova (1961)

Stage	Substage	Index	*Resheniya...*, 1955; Zonal subdivision for the Caucasus	I.G. Sazonova, 1958; Volga River basin	A.E. Glazunova, 1961; Ul'yanovsk region, Volga River basin
Aptian	lower	$Cr_1apt_1^4$	*Dufrenoya furcata*, *D. subfurcata*	Not dividable into zones	
		$Cr_1apt_1^3$	*Deshayesites dechyi*	*Deshayesites consobrinoides* (Sinz.), *D. deshayesi* (Leym.), *D. dechyi* (Papp), *Sinzovia trautscholdi* (Sinz.), *Corbula polita* Trautsch.	Horizon of uncoiled ammonites: *Ancyloceras gracilis* Sinz., *Ancyloceras* sp., *Deshayesites* sp., *Inoceramus* sp.;
					Deshayesites deshayesi Leym., *D. consobrinoides* Sinz., *Aconeceras trautscholdi* Sinz.
		$Cr_1apt_1^2$	*Deshayesites weissi*	*Deshayesites weissi* (Neum. et Uhl.), *Tropaeum bowerbanki* Sow., *Sinzovia trautscholdi* (Sinz.)	*Deshayesites weissi* Neum. et Uhl., *D. lavaschensis* Kas., *D. ssengillyensis* I. Sas., *Aconeceras trautscholdi* Sinz., *Tropaeum bowerbanki* Sow., *Ancyloceras* ex gr. *matheroni*
		$Cr_1apt_1^1$	*Tropaeum gillsi*, *Matheronites ridzewskyi*	In the south of the Volga River right bank area, south of Saratov; *Matheronites ridzewskyi* Kar.	

and other years. His valuable notes on Aptian ammonites were published a bit later (Sintsov, 1898, 1905). In 1898, he described in detail *Oppelia trautscholdi* Sinz. found near Simbirsk and Saratov, established two new species *Hoplites consobrinoides, H. subfissicostatus*, and reported about *H.* cf. *weissi* Neum. classified in the open nomenclature. Species *Ammonites deshayesi* (Leym.) pictured by Trautschold (1865, Plate III, nos. 16 a–b), as mentioned above, was suggested as a synonym of *H. consobrinoides*. Trautschold's collection is stored at the Timiryazev Agricultural Academy, however, we failed to find here the specimen in question. In 1905, Sintsov described heteromorphic ammonites from the Volga River basin: *Crioceras bowerbanki* J. de C. Sowerby, 1937, *Crioceras gracile* Sinzow, 1870, *Crioceras tuberculatum* Sinzow, 1870, *Crioceras tuberculatum* var. *graciloides* Sinzow, 1870, and *Hamites (?) eichwaldi* Jasykow. He also established a new species *Crioceras laticeps*.

In 1889, Nikitin correlated Aptian deposits of the Simbirsk area with analogous deposits of England and gave a brief characteristics to two ammonites: *Hoplites deshayesi* d'Orb. (Leym.) and *Amaltheus bicurvatus* Mich.

A purposeful study of lower Aptian deposits was resumed in the middle of the 20th century. Works by Sazonova (1954, 1956, 1957, 1958, 1961) and those in co-authorship with Sazonov (1967, 1991) lasted nearly 40 years. The work by Sazonova's *Lower Cretaceous Deposits of Central Areas of the Russian Platform* (1958) occupies an important place in this list that is far from being complete. This work contains description of particular sections in the Ul'yanovsk region (near Sengilei and Kremenki) and elucidates subdivision of these deposits and their correlation with coeval deposits of the Caucasus. Unfortunately, clays abundant in these sections caused numerous landslides that resulted in an erroneous interpretation of successive changes in ammonite assemblages. For instance, Sazonova (1958, p. 66) distinguished two ammonite zones in the Sengilei section and reported that *Deshayesites weissi* (Neum. et Uhl.), *Sinzovia trautscholdi* (Sinz.), *Tropaeum bowerbanki* Sow., *Ancyloceras* ex gr. *matheroni* (Orb.), and *D. ssengillyensis* I. Sason. are characteristic of the lower one, whereas various *Deshayesites* species and *Sinzovia trautscholdi* Sinz. were regarded as forms of the upper zone. Leaving aside some debatable identifications of the species, it is evident that the horizon with uncoiled ammonites occurs below the horizon, in which only monomorphic *Deshayesites* and *Sinzovia* were encountered. The lower assemblage was correlated with the *D. weissi* Zone and the upper one with the *D. dechyi* Zone (Table 1).

The *Matheronites ridzewskyi* Zone, shown in the Table below the two mentioned zones, is not substantiated by fauna from the Ul'yanovsk locality. As to the *Deshayesites weissi* Zone, it is most likely that it was distinguished near Ul'yanovsk in accord with presence of synonymous zone in the unified scheme elaborated for the Caucasus (*Resheniya...*, 1955).

The composition of ammonite assemblages of the *D. weissi* Zone and the overlying *D. dechyi* Zone engages attention. Along with representatives of genera *Sinzovia* and *Deshayesites*, ammonites in the lower of two mentioned zones are represented by heteromorphic *Ancyloceras* ex gr. *matheroni* (Orb.), *Crioceras* (?) *gracile* (Sinz.), *Tropaeum bowerbanki* (Sow.).

In the paleontological part of her work, Sazonova described several species of the *Deshayesites* genus (*D. deshayesi* Leym., *D. consobrinoides* Sinz., *D.* cf. *consobrinoides* Sinz., *D. volgensis* I. Sasonova, *D. weissi* Neum. et Uhl., *D. ssengillyensis* I. Sasonova, *D. evolvens* Lupov) and three species of the *Sinzovia* genus (*S. trautscholdi* Sinz., *S. luppovie* I. Sasonova, *S. saratoviensis* I. Sasonova). Among the described heteromorphic ammonites, there were only *Tropaeum bowerbanki* Sowerby (actually the spiral part of *Audouliceras*) and *Ancyloceras* cf. *matheroni* Orbigny (specimen no. 1 shown in her Plate XII does not belong to this genus and species). In their joint work (1991), Sazonova and Sazonov retained the zonal subdivision mentioned above.

Glazunova started her work in the Volga River middle courses in the 1960s (her works of 1961, 1967, 1968, and others). She collected fossils near Ul'yanovsk and village of Shilovka, i.e., northward from areas studied by Sazonova. "K.A. Kabanov from the Ul'yanovsk Regional Museum took part in some itineraries throughout the region" (Glazunova, 1973, p. 5). He granted some additional material for her study. Later, Kabanov's collection was purchased by the Paleontological Institute of the Russian Academy of Sciences (PIN RAS), and now it is stored here under no. 3390. Some specimens of the collection are exhibited in the Paleontological Museum.

Glazunova in her monograph *Paleontological Substantiation of Cretaceous Subdivisions in the Volga River Areas: The Lower Cretaceous* (1973) disagrees in many instances with inferences made by Sazonova. She calls into question two points. First, her idea of distinguishing the *Deshayesites weissi* Zone, since it is unlikely that the form pictured by Sazonova under this name represents the necessary species (Glazunova, 1973, p. 13). Second, the marker layer in the *D. deshayesi* Zone is known as the so-called Aptian plate, and ammonite assemblage of several *Deshayesites* and *Aconeceras* (= *Sinzovia* according to Sazonova) species is characteristic of the plate and beds below it. In contrast, ammonites found above the plate are of substantially different composition (as reported by Glazunova): in this assemblage, species of the *Deshayesites* genus coexist with abundant uncoiled ammonites. The term "horizon of uncoiled ammonites" was applied for this part of the section as early as in 1961.

Hence, the disagreement in stratigraphic position of the level with uncoiled ammonites is obvious: the level is below the *D. deshayesi* Zone according to Sazonova, whereas Glazunova placed it in the upper part of the *D. deshayesi* Zone above the Aptian plate. Thus, the succession of ammonite zones was inversely understood.

In 1961, Glazunova only noted that she failed to distinguish the *Deshayesites weissi* Zone of Sazonova in the section, but in 1973 she cast doubt on its existence. Glazunova (1973) described in detail a diverse assemblage of *Deshayesites* forms and established five new species of the genus: *D. variabilis, D. kabanovi, D. collevarus, D. lavaschensiformis*, and *D. imitator.* She also identified two new heteromorphic species of the *Australiceras* genus (*A. apticum, A. altum*) and additionally distinguished in Sintsov's collection three species and one variety: *Australiceras rossicus, A. solidum, A. jasykowi, A. simbirskense* var. *sitschevkensis.*

Though the Aptian stratigraphy for the Volga River region has been elaborated in detail, the latest data made it possible to reconsider and supplement the available stratigraphic schemes (Baraboshkin, 1998; Baraboshkin *et al.*, 1999; Mikhailova and Baraboshkin, 2001). Biostratigraphic schemes for this region, like for many other regions of the world (Casey, 1961; Casey *et al.*, 1998; Kemper, 1995, and others), have been worked out on the basis of ammonites of the family Deshayesitidae. In the last decade, lower Aptian arenaceous–argillaceous rocks of the Volga River region were found to contain a rich assemblage of heteromorphic ammonites of the family Ancyloceratidae poorly known before.

Numerous specimens found by Shumilkin, Uspenskii, Efimov, Krivosheev (Ul'yanovsk) and by us, as well as the restudy of specimens collected by K.A. Kabanov and G.K. Kabanov (PIN RAS) showed an extreme diversity of ancyloceratids from the Volga River region. Among them, we identified for the first time many genera, which were unknown in the study area. The revealed succession of heteromorphic genera and species in the section allowed us to propose a new biostratigraphic scheme for the lower Aptian (Mikhailova and Baraboshkin, 2001). Accordingly, we feel it necessary to present here brief information on the lower Aptian biostratigraphy in the Ul'yanovsk–Saratov area of the Volga River region.

The lower Aptian section traceable between the village of Kremenki and Ul'yanovsk is most complete, having the best paleontological characterization (Baraboshkin, 1998; Baraboshkin *et al.*, 1999). The section can be regarded as the reference one for the lower Aptian of the region.

STRUCTURE OF THE REFERENCE SECTION

The section begins near the village of Kremenki (Fig. 1); its lower member is composed of silty clays with interlayers of loose sandstones with siderite concretions. Clays are gray to brownish, bioturbated, containing rare casts and imprints of *Cymbula* aff. *nuda* (Keys.) and *Nucula* sp. An interlayer of bioturbated sandstones, 0.4 m thick, occurs at the top. At the base, the member is bounded by a horizon of giant (up to 1 × 5 m) carbonate concretions. The upper boundary is represented by the "softground" surface and pierced by fucoids of the *Scolithos* type, which are filled with overlying clays. Species *Oxyteuthis lahuseni* (Pavl.), *O. barremicus* Glas., *Oxyteuthis* sp., and fragments of *Cucullaea golowkinskii* (Sinz) were encountered at the

base of sandstones. As for rare belemnites found in the section, their rostra are highly weathered and replaced by gypsum. Species *Oxyteuthis lahuseni* (Pavl.), *O.* aff. *germanica* Stoll, *O.* sp. were found directly below landslides on the sloping beach. The member is 21.7 m thick. It crowns the section of Barremian deposits, the age of which is confirmed by the presence of the reverse magnetic polarity zone (an analog of Chron M0) and by appearance of Aptian dinoflagellate form *Pseudoceratium eisenackii* at the base of the overlying member (Baraboshkin *et al.*, 1999). The following units are distinguished above the latter:

Member I: rhythmic alternation of gray–brown, loose sandstone (0.03–0.1 m), dark gray, clayey silt (0.8–2.0 m), and black clay. Rocks are bioturbated. Clay encloses abundant weathered marcasite (pyrite) concretions. As a rule, sandstone interlayers bear carbonate concretions, and their roof usually corresponds to the "softground" surface. Upper parts of the rhythms are brown, colored with iron oxides. The member includes three rhythms, 10.2 m thick in total.

Member II: rhythmic alternation of dark gray, silty, bioturbated clay (2–5 m) and brown, loose, glauconite–quartz sandstone (0.2–0.5 m). Clays bear weathered marcasite (pyrite) concretions, whereas carbonate concretions are characteristic of sandstones. "Softground" surfaces are usually encountered in the roof of sandstone beds. The member includes four rhythms. The total thickness is 22–23 m. The ammonite species *Deshayesites* cf. *tenuicostatus* (von Koenen, 1902), as well as bivalves *Arctica* (?) sp. and *Cymbula nuda* (Keys.) were encountered in landslide from the upper part of the member.

Units described further are exposed near a new bridge in Ul'yanovsk and continue the section.

Member III: rhythmic alternation of green–brown loose glauconite–quartz sandstone (0.2–0.5 m), dark gray clay (0.2–3.0 m), gray banded bioturbated clay (1.5–2.0 m), and silty clay with siderite concretions. "Softground" surfaces occur at the base of each sand interlayer. The member includes three rhythms, the total thickness of which is 7.8 m. The roof of the upper rhythm is eroded. The ammonite species *Deshayesites* cf. *tenuicostatus* (von Koenen, 1902), bivalves *Cymbula nuda* (Keys.) and *Neocomiceramus volgensis* (Glas.), and abundant serpulids *Ditrupa notabile* (Eichw.) were encountered in the member.

Member IV: black combustible shale with large, closely spaced carbonate concretions near the base and with smaller concretions in the upper part; the member exhibits fine alternation of light and black laminae (1–5 mm thick). Large wood fragments, shell detritus, and small phosphate concretions occur at the member base. One or two erosion surfaces are distinguishable in the basal part. Abundant flattened ammonites, aptychi, and fish scales are confined to bedding planes. Some of the latter are almost completely covered with embryonic ammonite shells. These peculiarities, along with a high C_{org} content (up to 6–8%), are indicative of anoxic conditions during sedimentation. Ammonites encountered in the member represent taxa *Deshayesites gracilis* Casey, 1964, *D. volgensis* Sasonova, 1958, *D. forbesi* Casey, 1961, *D. consobrinoides* (Sinzow, 1898), *D. saxbyi* Casey, 1964, *D.* aff. *vectensis* Spath, 1930, *D.* sp., *Paradeshayesites imitator* (Glasunova, 1968), *Obsoleticeras levigatum* (Bogdanova, 1991), and *Sinzovia trautscholdi* (Sinzow, 1870). These forms associate with smaller heteromorphs *Volgoceratoides schilovkensis* I. Michailova et Baraboshkin, sp. nov., *Koeneniceras tenuiplicatum* (von Koenen, 1902), *K. rareplicatum* I. Michailova et Baraboshkin, sp. nov., and with bivalves *Cymbula* sp. and *Phacoides borealis.* It was this level from which Glazunova (1973) had reported findings of *Deshayesites deshayesi.* The member is 3.8–4 m thick.

Member V: homogeneous, dark gray clay with disseminated shell detritus at the base; collected ammonites are *Deshayesites multicostatus* Swinnerton, 1935, *D. consobrinoides* (Sinzow, 1898), *D.* sp., *Paradeshayesites ssengillyensis* (Sasonova, 1958), *P. callidiscus* (Casey, 1961), *P. topleyi* (Spath, 1930), *P. similis* (Bogdanova, 1991), *P. imitator* (Glasunova, 1968), heteromorphic ammonites: *Ancyloceras matheronianum* d'Orbigny, 1842, *Lithancylus* aff. *grandis* (J. de C. Sowerby, 1829), *L. glebi* I. Michailova et Baraboshkin, 2001, *L. grandis* (J. de C. Sowerby, 1829) *L. igori* I. Michailova et Baraboshkin, 2001, *L. russiensis* I. Michailova et Baraboshkin, *L. tirolensiformis* I. Michailova et Baraboshkin, 2001. The thickness is 3–3.2 m.

Member VI: dark gray, bioturbated, silty clay with rare extended lenticular interlayers rich in glauconite–quartz sandy material; organic remains are abundant separated valves *Arctica anglica* (Woods), *Cymbula gardneri* (J. Nikit.), *Modiolus* sp., *Thetironia* sp., *Panopea neocomiensis* (Leym.), *Corbula* sp., *Neocomiceramus volgensis* (Glas.), and *N. borealis* (Glas.). Two horizons of carbonate concretions and disseminated phosphorite are recorded. Ammonite taxa encountered in clay and concretions are *Deshayesites* aff. *rarecostatus* Bogdanova, Kvantaliani, and Scharikadze, 1979, *D.* sp., *Paradeshayesites ssengillyensis* (Sasonova, 1958), *Cheloniceras* ex gr. *cornuelianum* (d'Orbigny, 1841), and nautiluses *Cymatoceras* aff. *bifurcatum* (Ooster, 1858), *C. karakaschi* Shimansky, 1975, *C.* cf. *karakaschi* Shimansky, 1975. Heteromorphic ammonites are characteristic of two levels. The lower one yields species *Proaustraliceras tuberculatum* (Sinzow, 1870), *P. rossicum* (Glasunova, 1973), *P. laticeps* (Sinzow, 1905), *P.* sp., *Pseudoaustraliceras pavlowi* (Vassilevsky, 1908), and *Toxoceratoides* sp. Forms identified at the upper level are *Audouliceras renauxianum* (d'Orbigny, 1842), *Toxoceratoides royerianus* (d'Orbigny, 1842), *T.* ex gr. *royerianus* (d'Orbigny, 1842), and *T.* sp. This member is known as the "horizon of uncoiled ammonites." Its thickness is 4 m.

Member VII: rhythmic alternation of gray clayey silt (0.2 m) and dark gray glauconite-bearing clay (0.2–0.3 m) with shell detritus and fragments of *Cymbula nuda* (Keys.) and *Neocomiceramus borealis* (Glas.). At the base, there are large flat siderite concretions with large ammonites *Tropaeum (Tropaeum) bowerbanki* (J. de C. Sow.) up to 80 cm in diameter and with clasts of inocerams *Neocomiceramus* cf. *borealis* (Glas.). A "softground" surface is developed at the member roof. The thickness is 1.6–1.8 m.

Member VIII begins the middle Aptian sequence. It is represented by dark gray bioturbated clay with shell detritus and several horizons of carbonate and marcasite concretions, septaria included. At the base, there is a thin silt interlayer, and the erosion surface marks the member top. Ammonites collected from the unit are *Tonohamites* sp., *Aconeceras nisum* (d'Orb.), *Nuculana lineata* (Sow.), *N.* sp., *Cymbula gardneri* (J. Nikit.), *Modiolus* cf. *subsimplex* (d'Orb.), *M. reversus* (J. Sow.), *Neocomicermus* cf. *borealis* (Glas.) *Arctica sedgwicki sedgwicki* (Walker), *Venilicardia (V.) protensa* (Woods), *V. (V.)* sp., *Panopea neocomiensis* (Leym.), and *Dentalium?* sp. The thickness is 7 m.

A similar section near Sengilei (Fig. 1) is supplemented with data of I.A. Shumilkin. It is analogous to that located near Ul'yanovsk being less complete in its lower part and, by contrast, more complete in the uppermost lower Aptian interval. We traced the described deposits along the Volga River right bank far to Saratov (Fig. 2).

LOWER APTIAN AMMONITE ASSEMBLAGES FROM THE UL'YANOVSK–SARATOV AREA

Ammonites have not been found in Member I, and its stratigraphic position is established on the basis of dinocyst assemblage and paleomagnetic data (Baraboshkin *et al.*, 1999). First rare ammonites, identified previously as *Deshayesites forbesi* Casey (Baraboshkin, 1998), appear in the upper part of Member II. Revising them together with T.N. Bogdanova, we came to the conclusion that they should be referred to as *Deshayesites* cf. *tenuicostatus* (von Koenen, 1902). As is shown by Bogdanova and Mikhailova (1999), the species are affiliated to the *Deshayesites* but not to *Prodeshayesites* genus (according to Casey, 1960–1980, 1961) that complies with opinion of Kemper (1995) as well. The assemblage from this stratigraphic level is more diverse near Saratov, where the arenaceous facies yield such forms as *Deshayesites tenuicostatus* (von Koenen, 1902, Plate I, fig. 3), *D.* ex gr. *tenuicostatus* (von Koenen, 1902), *D. bodei* (von Koenen, 1902), and *D.* aff. *bodei* (von Koenen, 1902, Plate I, fig. 2), and *D.* sp. (Fig. 3).

In Member III of the Ul'yanovsk section, a single specimen of *Deshayesites* cf. *tenuicostatus* was encountered (von Koenen, 1902). The upper part of the member is eroded in the Ul'yanovsk area, while in the more complete sections near Khvalynsk, we encountered *Deshayesites volgensis* Sasonova, 1958 and *D.* sp. directly below combustible shales of Member IV.

Member IV of combustible shales (the "Aptian plate" level) yields abundant ammonites *Deshayesites gracilis* (Casey, 1964), Plate II, fig. 3; *D. volgensis* (Sasonova, 1958), Plate III, fig. 1; *D. forbesi* (Casey, 1961), Plate III, fig. 2; *D. consobrinoides* (Sinzow, 1898), *D. saxbyi*, (Casey,1964), Plate IV, fig. 2; *D.* aff. *vectensis* (Spath, 1930); *D.* sp., *Paradeshayesites imitator* (Glasunova, 1968); *Obsoleticeras levigatum* (Bogdanova, 1991), Plate II, fig. 1; and *Sinzovia trautscholdi* (Sinzow, 1870), Plate V, fig. 4. Among forms appearing here for the first time, there are small heteromorphic ammonites *Volgoceratoides schilovkensis* I. Michailova et Baraboshkin, sp. nov., Plate VI, figs. 8–11; *Koeneniceras tenuiplicatum* (von Koenen, 1902), Plate VI, figs. 2–5; and *K. rareplicatum* I. Michailova et Baraboshkin, sp. nov., Plate VI, fig. 12.

Forms identified in Member V are *Deshayesites multicostatus* Swinnerton, 1935; *D. consobrinoides* (Sinzow, 1898); *D.* sp. and *Paradeshayesites ssengillyensis* (Sasonova, 1958); *P. callidiscus* (Casey, 1961), Plate I, fig. 1; *P. topleyi* (Spath, 1930); *P. similis* (Bogdanova, 1991), Plate IV, fig. 1; *P. imitator* (Glasunova, 1968), Plate III, fig. 3; and abundant heteromorphic ammonites: *Ancyloceras matheronianum* d'Orbigny, 1842, Plate VI, fig. 7; *Lithancylus* aff. *grandis* (J. de C. Sowerby, 1829); *L. glebi* I. Michailova et Baraboshkin, 2001; *L. grandis* (J. de C. Sowerby, 1829); *L. igori* I. Michailova et Baraboshkin, 2001; *L. russiensis* I. Michailova et Baraboshkin, 2001, Plate V, fig. 2; *L. tirolensiformis* I. Michailova et Baraboshkin, 2001. These forms were found mainly in the upper part of the member. It was believed previously (Baraboshkin *et al.*, 1999) that a stratigraphic hiatus corresponds to this interval in the Ul'yanovsk section. The *Ancyloceras matheronianum* species, found recently by Shumilkin and Uspenskii, confirms presence of the interval with *Ancyloceras* in this section, although it is half as thick as the Sengilei section.

Member VI yields both the monomorphic deshayesitids and heteromorphic ancyloceratids. The Deshayesitidae assemblage is homogeneous for the whole member and includes *Deshayesites* aff. *rarecostatus* Bogdanova, Kvantaliani, and Scharikadze, 1979, *D.* sp., and *Paradeshayesites ssengillyensis* (Sasonova, 1958). In addition, among the encountered forms there were *Cheloniceras* ex gr. *cornuelianum* (d'Orbigny, 1841), Plate IV, fig. 3; *Cymatoceras* aff. *bifurcatum* (Ooster, 1858), Plate II, fig. 2; *C. karakaschi* Shimansky, 1975; and *C.* cf. *karakaschi* Shimansky, 1975.

Representatives of the family Ancyloceratidae constitute two assemblages. The lower consists of *Proaustraliceras tuberculatum* (Sinzow, 1870), Plate V, fig. 1 (= *Ancyloceras simbirskensis* Jasykow sensu Laguzen, 1974, p. 70, Plate VIII, fig. 2 = *Australiceras simbirskense*, Sinzow, 1872, sensu Glazunova, 1973, Bara-

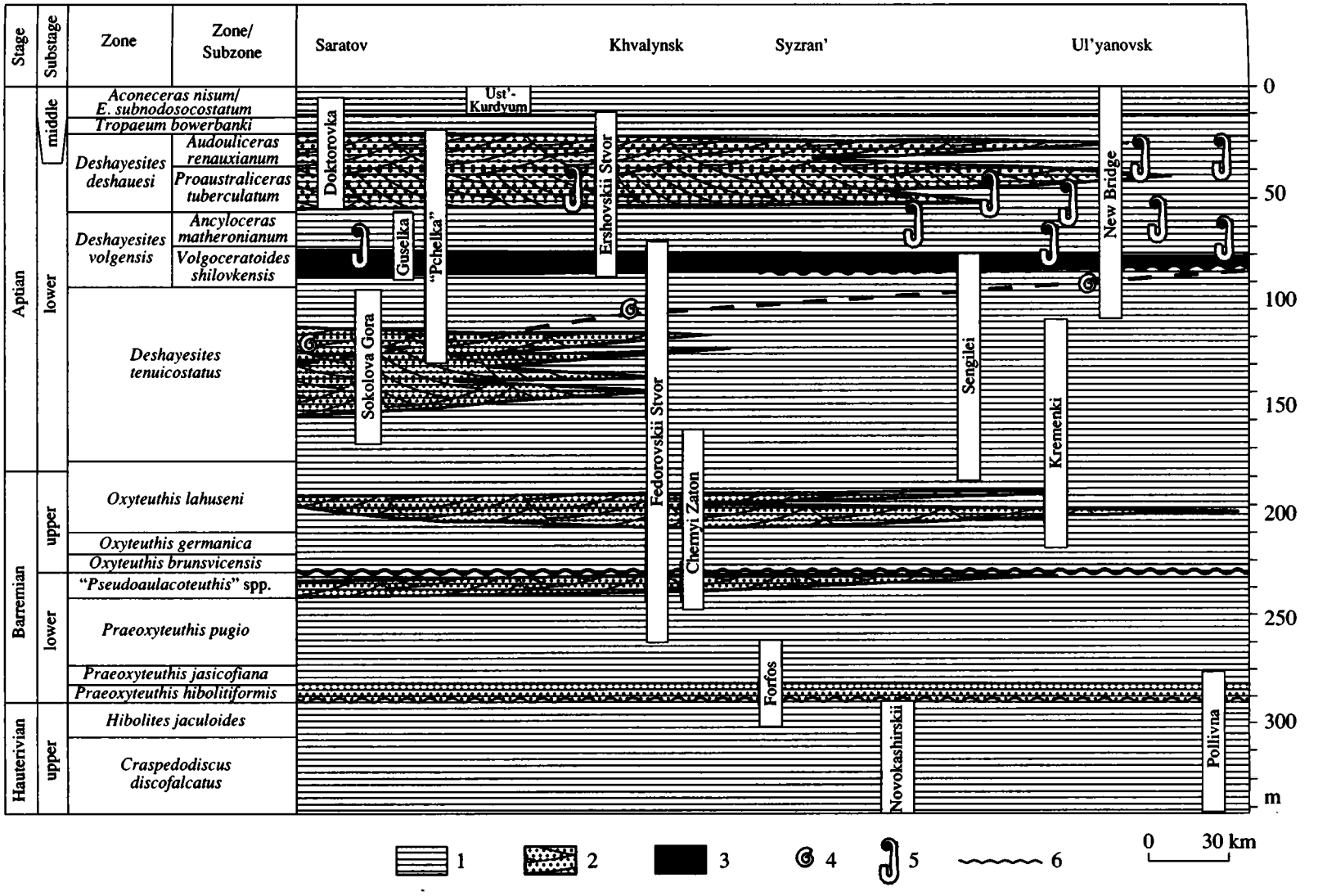


Fig. 2. Facies relationships and localities of lower Aptian sections along the Volga River right bank between Ul'yanovsk to Saratov: (1) clay; (2) obliquely laminated sand; (3) combustible shale; (4) findings of *Deshayesites tenuicostatus*; (5) findings of heteromorphic ammonites; (6) most significant stratigraphic hiatuses.

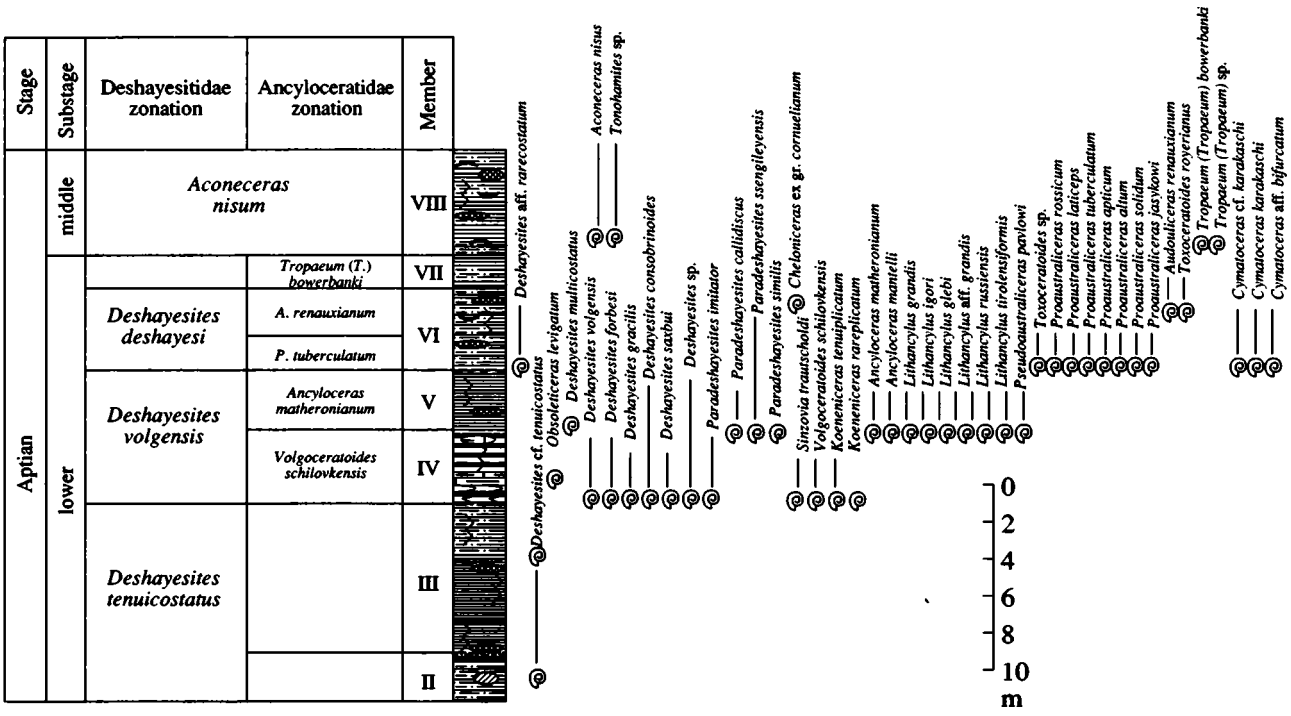


Fig. 3. Stratigraphic ranges of cephalopod mollusks in the lower Aptian section, the Ul'yanovsk region (symbols as in Fig. 1).

boshkin, 1998); *"A." rossicum* (Sasonova, non Casey; = *"Crioceras gracile,"* Sintsov, 1905); *"A."* sp., *Proaustraliceras laticeps* (Sinzow, 1905); *Pseudoaustraliceras pavlowi* (Vassilevsky, 1908), Plate V, fig. 3; and *Toxoceratoides* sp.

The upper assemblage is less diverse, consisting of *Audouliceras renauxianum* (d'Orbigny, 1842), Plate VI, fig. 1; *Toxoceratoides royerianus* (d'Orbigny, 1842), Plate VI, fig. 6; *T.* ex gr. *royerianus* (d'Orbigny, 1842); and *T.* sp.

It is quite likely that *"Australiceras" apticum* and *"A." altum*, which were identified by Glazunova (1973), and *"A." rossicum*, *"A." solidum*, and *"A." jasykowi* and listed based on data by Sintsov (1905) and which should be attributed to the genus *Proaustraliceras*, can characterize the lower part of Member VI. This inference needs however a further verification.

Member VI is more sandy southward, and in the Saratov area it is represented by obliquely laminated sand and sandstone beds, which yield only rare and large *Deshayesites* forms indicative of a strict facies control over the assemblage compositions (Fig. 2).

Member VII crowning the lower Aptian sequence bears rare giant ammonites *Tropaeum* (*Tropaeum*) *bowerbanki* J. de C. Sowerby, 1837, Plate III, fig. 4, and *T.* sp. It is also more sandy in southern sections.

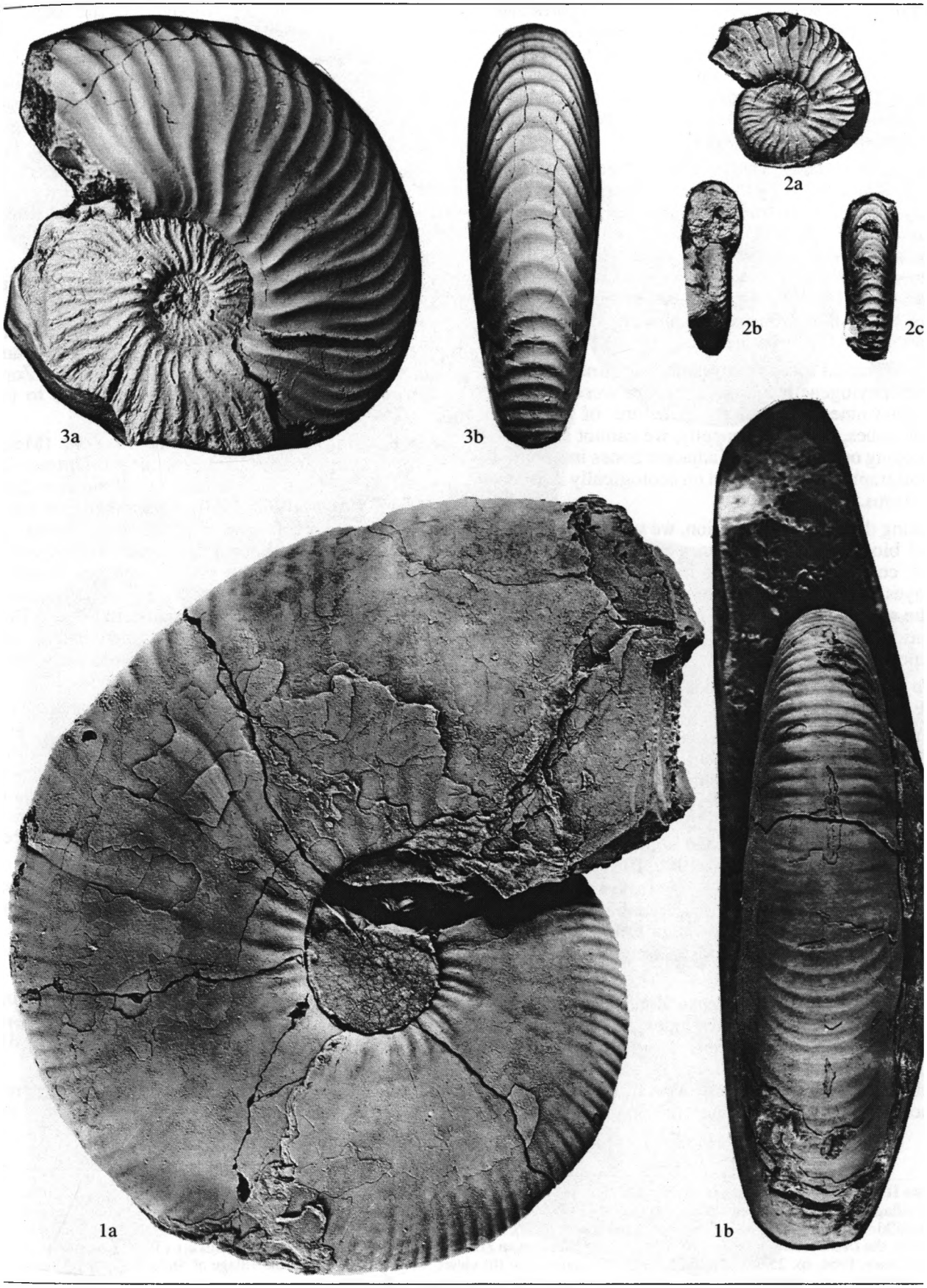
BIOSTRATIGRAPHIC SCHEME FOR THE LOWER APTIAN IN THE VOLGA RIVER MIDDLE COURSES

Owing to new comprehensive data on the distribution of ammonites in sections of the Volga River middle courses, we got a possibility to elaborate two parallel biostratigraphic schemes based on the phylogenetic evolution of representatives from families Deshayesitidae and Ancyloceratidae. For some reasons, such schemes are more preferential compared to synthetic schemes commonly used to subdivide the lower Aptian (Casey, 1960–1980, 1961; Casey *et al.*, 1998; Kemper, 1967, 1995, and others).

Various ammonite groups lived in different ecological (and bathymetric) environments (Westermann, 1990; Bengtson and Kakabadze, 1999, and others) that

PLATES I–VI. Zonal and characteristic cephalopod species from lower Aptian deposits of the Volga River middle courses (natural dimensions if otherwise not specified).

Plate I. (1) *Paradeshayesites callidiscus* (Casey, 1961), sp. 24/96 MPG MGU, side (a) and aperture (b) views, Ul'yanovsk region, the village of Shilovka. lower Aptian, the *Deshayesites volgensis/Ancyloceras matheronianum* Zone (collected by V.M. Efimov). (2) *Deshayesites* aff. *bodei* (von Koenen, 1902), sp. 21/96 MPG MGU, side (a), aperture (b), and ventral (c) views, Saratov, Sokolova Gora, lower Aptian, the *Deshayesites tenuicostatus* Zone (collected by Baraboshkin). (3) *Deshayesites tenuicostatus* (von Koenen, 1902), sp. 22/96 MPG MGU, side (a) and ventral (b) views, the same locality and stratigraphic range (collected by Baraboshkin).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determined their strict facies confinement (in particular, Kakabadze, 1981). This is readily apparent from comparison of ammonite assemblages characterizing the deep-sea sections of the Ul'yanovsk region, which are rich in representatives of Ancyloceratidae, and shallow-water sections of the Saratov area, where heteromorphic ammonites are extremely rare.

The abundance of Deshayesitidae and Ancyloceratidae was controlled by the water temperature as well: *Deshayesites, Proaustraliceras, Ancyloceras, Volgoceratoides*, and *Koeneniceras* forms populated only the northern hemisphere, whereas *Audouliceras, Tropaeum*, and *Lithancylus* taxa were found in both hemispheres (Day, 1969; Klinger and Kennedy, 1977; Kakabadze, 1981; Bengtson and Kakabadze, 1999), where they populated two separate areas (Fig. 4).

The aforesaid means that spatial and time ranges of different phylogenetic ammonite groups were different. This determines the provincial nature of biostratigraphic zones, and, consequently, we cannot avoid the overlapping or gaps between adjacent zones in a unified biostratigraphic scheme based on ecologically different zonal forms.

Taking this into consideration, we suggest two interrelated biostratigraphic schemes for the Volga River middle courses: the first one based on evolution of Deshayesitidae characteristic of shallow-water deposits and the second scheme based on distribution of Ancyloceratidae characterizing deep-water pelagic sequences.

We regard the Deshayesitidae zonation as follows (Table 2):

(1) The *Deshayesites tenuicostatus* Zone (Member II and the lower part of Member III) bears the ammonite assemblage similar to that from the *tenuicostatus* and *bodei* zones of northern Germany (Kemper, 1967, 1995), but we failed to subdivide it into two subunits. The zone can be also correlated with the *fissicostatus* Zone of England (Casey, 1960–1980, 1961), but its correlation with zonal schemes of Turkmenistan and Northern Caucasus (Bogdanova and Mikhailova, 1999), and with that of southeastern France (Delanoy, 1995) is impossible, because the analogous index species are unknown.

(2) The *Deshayesites volgensis* Zone (the upper part of Member III–Member V), the index species of which is very similar to and has been previously identified with *Deshayesites forbesi* Casey, 1961, from England (Baraboshkin, 1998; Baraboshkin *et al.*, 1999); it is not inconceivable that being further studied the last species could be included into the synonymy of the *volgensis* form described by Sazonova in 1958.

(3) The *Deshayesites deshayesi* Zone (Member VI) is distinguished based on its ammonite assemblage, though the corresponding index species has not been encountered in the studied sections. Accordingly, nomenclature can be likely changed in the future. The zone is correlative with synonymous zones of Turkmenistan, Northern Caucasus (Bogdanova and Mikhailova, 1999), England (Casey, 1960–1980, 1961; Casey *et al.*, 1998), Germany (Kemper, 1995), and France (Delanoy, 1995).

Characterization of Ancyloceratidae zonation is as follows (Table 2):

(1) The *Volgoceratoides schilovkensis* Zone (Member IV); representatives of the *Volgoceratoides* and *Koeneniceras* genera are known from the *weissi* Zone of Germany (Koenen, 1902) that corresponds to the *forbesi* Zone of England.

(2) The *Ancyloceras matheronianum* Zone (Member V) bearing ammonite analogous to *Lithancylus grandis* known from the *Deshayesites deshayesi* Zone of England (Casey, 1960–1980). According to the latest data, *Lithancylus* cf. *grandis* and *Ancyloceras* cf. *matheronianum* are known from the *Deshayesites annelidus* Subzone of the *forbesi* Zone of England (Casey *et al.*, 1998), and *Ancyloceras matheronianum* is found in the *weissi* Zone of southeastern France (Delanoy, 1995). These data confirm that mentioned ammonite taxa occur near the boundary between the *forbesi* and *deshayesi* zones.

(3) The *Proaustraliceras tuberculatum* Zone (the lower part of Member VII); representatives of the genus *Proaustraliceras* are known from the upper part of the *Deshayesites grandis* Subzone (the *deshayesi* Zone) of England and from the lower part of the *Tropaeum bowerbanki* Zone (Casey *et al.*, 1998). They are characteristic of lower Aptian deposits in other regions as well (Kakabadze, 1981).

(4) The *Audouliceras renauxianum* Zone (the upper part of Member VII); the index taxon is known from lower Aptian deposits of southeastern France (Thomel, 1964; Delanoy, 1995) and the Caucasus (Kakabadze, 1981), though without the exact zonal position.

(5) The *Tropaeum* (*Tropaeum*) *bowerbanki* Zone (Member VIII); its index species characterizes the *bowerbanki* Zone of England (Casey, 1960–1980, 1961; Casey *et al.*, 1998).

As it is evident, the Ancyloceratidae zonation in relatively deep-water facies of the Ul'yanovsk area of the

Plate II. (1) *Obsoleticeras levigatum* (Bogdanova, 1991), sp. 20/96 MPG MGU, side (a) and ventral (b) views, Ul'yanovsk region, the village of Kriushi, Lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by Uspenskii and Shumilkin). (2) *Cymatoceras* aff. *bifurcatum* (Ooster, 1858), sp. 28/96 MPG MGU, side view, Ul'yanovsk region, Sengilei, lower Aptian, the *Deshayesites deshayesi/Audouliceras renauxianum* Zone (collected by Uspenskii and Shumilkin). (3) *Deshayesites gracilis* Casey, 1964, sp. 25/96 MPG MGU, side (a) and ventral (b) views, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by Efimov).

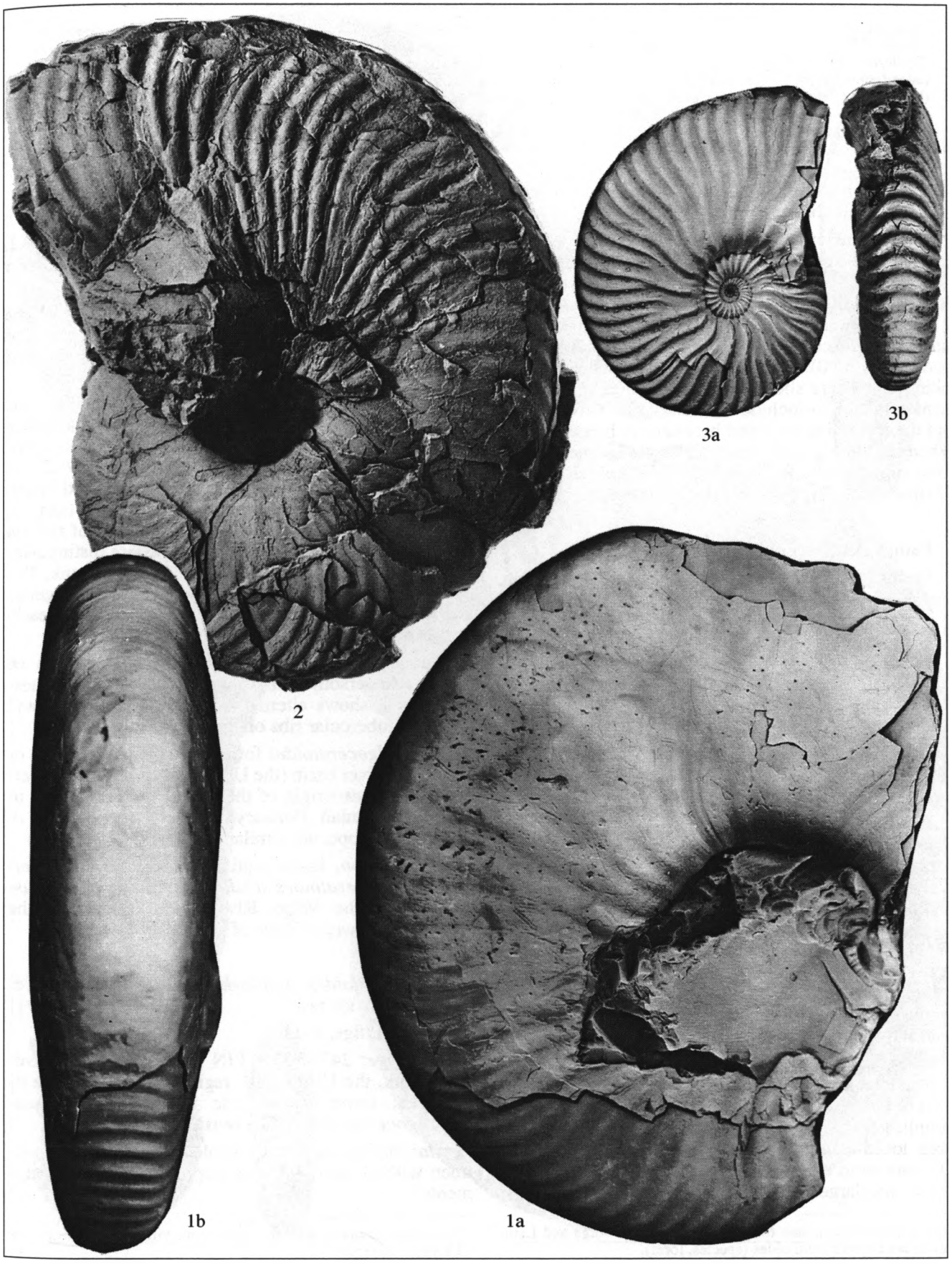
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Plate III. (1) *Deshayesites volgensis* Sasonova, 1958, sp. 30/96 MPG MGU, side (a) and ventral (b) views, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by G.K. Kabanov). (2) *Deshayesites forbesi* Casey, 1961, sp. 2291/3390 PIN RAS, side (a) and aperture (b) views, the same locality and stratigraphic range (collected by K.A. Kabanov). (3) *Paradeshayesites imitator* (Glasunova, 1968), sp. 29/96 MPG MGU, side (a) and aperture (b) views, the same locality, lower Aptian, the *Deshayesites volgensis/Ancyloceras matheronianum* Zone (collected by Efimov). (4) *Tropaeum* (*Tropaeum*) *bowerbanki* J. de C. Sowerby, 1837, sp. 13/96 MPG MGU, side view, magnification ×0.39, Ul'yanovsk, lower Aptian, the *Tropaeum bowerbanki* Zone (collected by Baraboshkin).

Volga River basin is more detailed and more easily recognizable than that of Deshayesitidae.

In conclusion, we present description of new generic and species ammonite taxa. Specimens are stored at the Museum of Physical Geography, Moscow State University (MPG MGU, Collection no. 96), and at the Paleontological Museum, Paleontological Institute, Russian Academy of Sciences (PM PIN RAS, Collection no. 3390). Size parameters of ammonoids and their symbols are shown in Fig. 5. Tables of measurement results also include data on angles between ribs and the trunk line (α_1) and between rib branches (α_3). When describing lobe lines, we used nomenclature of lobes suggested by Ruzhentsev: (V) ventral; (L) lateral; (U) umbilical; (I) internal lateral; (D) dorsal.

Family Ancyloceratidae Gill, 1971

Genus *Volgoceratoides* I. Michailova et Baraboshkin, gen. nov.[1]

Ancyloceras (pars): Koenen, 1902, p. 331.

Toxoceratoides (pars): Klinger and Kennedy, p. 305.

Toxoceratoides (pars): Aguirre Urreta, 1986, p. 295.

Type species: *Volgoceratoides schilovkensis* I. Michailova et Baraboshkin, sp. nov., lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone, the Volga River right bank, the Ul'yanovsk region, the village of Shilovka.

Diagnosis. Small hamulicones with arched shells (widely uncoiled spiral) at the early stage and with a trunk and hamulus at the middle and late stages, respectively. Their first whorl and initial chamber are unknown. The section height is less than 1 mm at the beginning of the arch of two specimens. Hence, it is logic to assume that the arching began almost immediately after the first whorl. The living chamber spans half of the trunk and the hamulus. Trunk is sculptured with simple ribs having two rows of tubercles (the lateral and ventral rows); on the hamulus ribs are bifid and intercalating.

The lobe line (Fig. 6) is simple due to small dimensions of *Volgoceratoides* shells. The bifid ventral lobe is complicated by two lateral scallops. The remaining three lobes—umbilical (U), internal (I), and dorsal (D)—are trifid and almost symmetrical. The umbilical lobe is the largest (wide and deep) one, whereas the internal lobe is the smallest. Saddles are bifid, the external saddle (V/U) is asymmetrical.

Composition. *Volgoceratoides schilovkensis* I. Michailova et Baraboshkin, sp. nov., *V. biplicatum* (von Koenen, 1902).

Comparison. The genus differs from *Toxoceratoides* and analogous *Helicancylus* genera in the absence of thickened three-tubercle ribs and in presence of bifid ribs on the hamulus.

When compared with other smaller hamulicones, the genus is the most similar to the upper Barremian genus *Hamulinites*, especially to the group "*Eoleptoceras* (*Tzankoviceras*) *tzankovi* Manolov, 1962" (= *Hamulinites parvulus* according to Vasicek and Wiedmann, 1994). In distinction from mentioned taxa, its smaller hamuli have a peculiar ramification of ribs on their bends, and two rows of tubercles are distinguishable on ribs at the end of spiral and on the trunk. The last of mentioned features is characteristic of the genus *Karstenites*, shells of which are convolute at early stages but lack hamuli.

The genus *Hamiticeras* (especially its species *H. pilsbryi* Anderson) is also similar to the genus in question, but it shows alternation of thick tubercular and thin non-tubercular ribs on the trunk.

The *Volgoceratoides* forms occurring in sections of the Volga River basin (the Ul'yanovsk region) and Germany elucidate origin of the genus: representatives of upper Barremian *Parancyloceras*, the endemics of Northern Europe, are similar in morphology to them.

Distribution, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone of the Russian plate (the Volga River middle courses), the *Deshayesites weissi* Zone of Germany.

Volgoceratoides schilovkensis I. Michailova et Baraboshkin, sp. nov.[2]

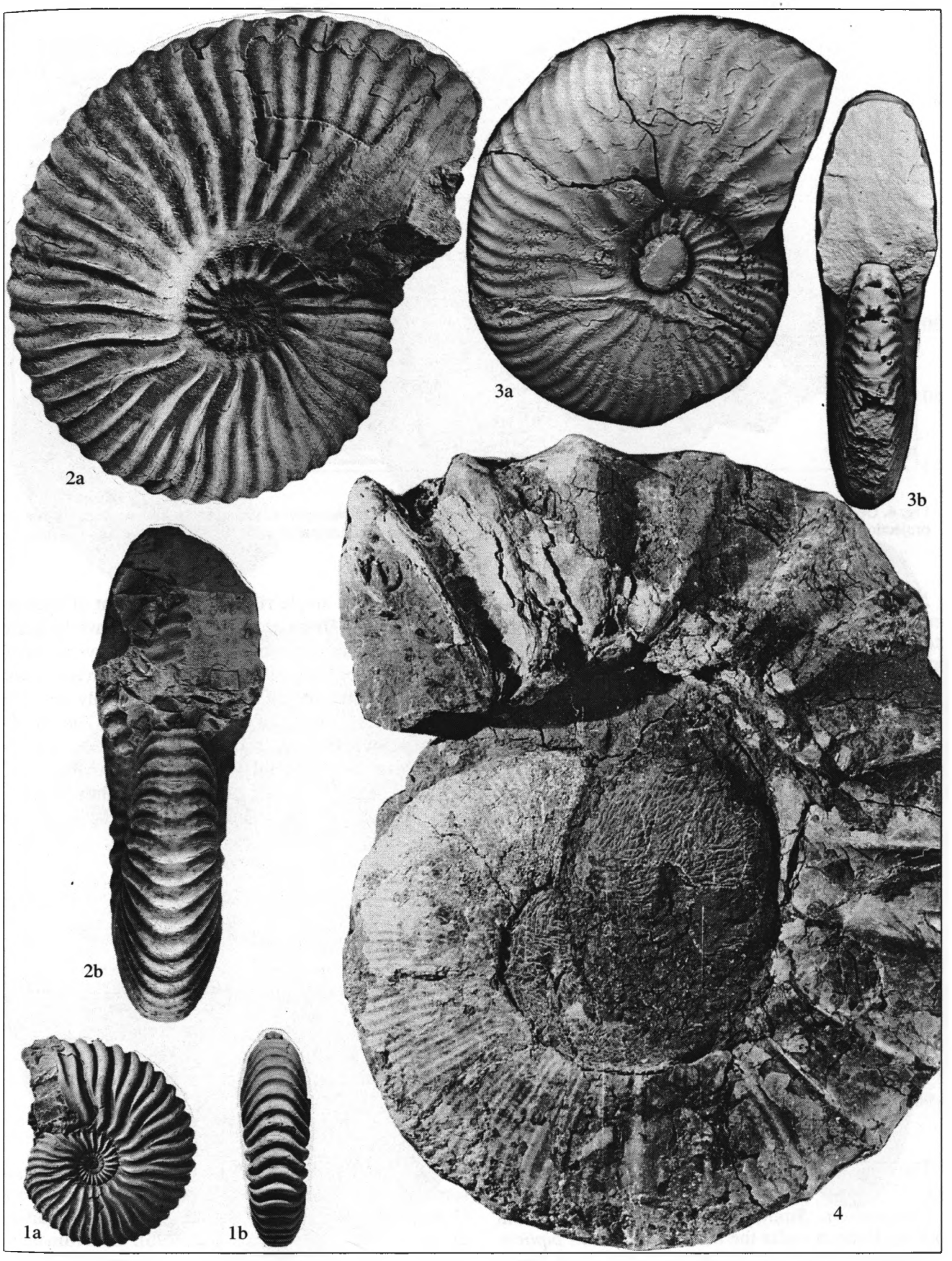
Plate VI, figs. 8–11

Holotype: 2478/3390, PIN RAS, the Volga River right bank, the Ul'yanovsk region near the village of Shilovka; lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone.

Material: seven almost whole specimens, one specimen without hamulus, one imprint and several fragments.

[1] The genus name means the eponymous region Volga and Latin terms keras (horn) and oides (species, form).

[2] The species name is after the eponymous village of Shilovka, the Ul'yanovsk region.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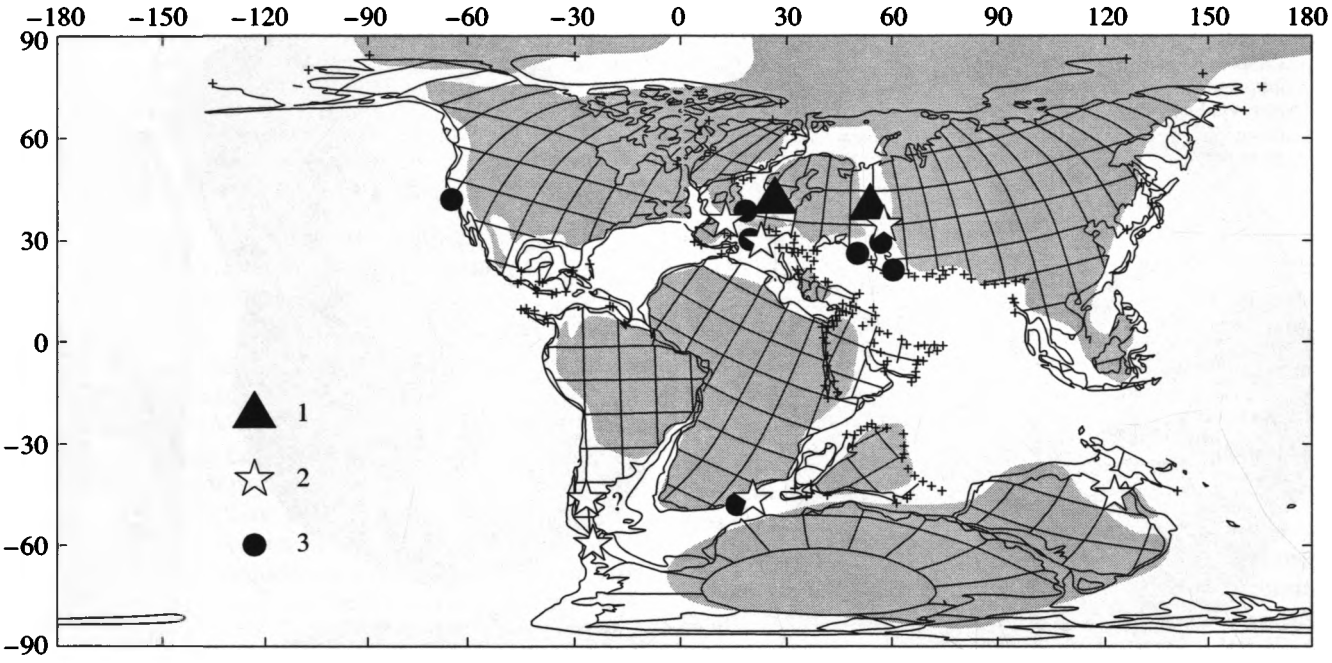

Fig. 4. Geographic distribution of genera *Volgoceratoides* and *Koeneniceras* (1), *Lithancylus* (2), and *Audouliceras* (3). Mercator projection and position of continents after Smith *et al.*, (1981); configuration of continents after Smith *et al.*, 1994 and authors' data.

Description. Shell is small (about 300 mm high), symmetrical, arched, terminated by hamulus, and characterizing three morphogenetic stages. The early stage corresponds to the arched shell 10–15 mm high (wide uncoiled spiral); short trunk appears at the middle stage and hamulus at the late stage. The living chamber occupies the middle part of the trunk and the hamulus. The cross-section is rounded–hexagonal with the maximum flattening on ventral side. We did not observe protoconchs and first whorls.

Sculpturing is clear, when trunk is more than 2 mm high. On the uncoiled spiral part, there are vague, widely spaced, single ribs. From the point of appearance and through the entire trunk, shells have frequent single ribs with two, ventral and lateral rows of small tubercles. Ribs are weakened between tubercles on the ventral side, but they do not disappear. Only near the hamulus bend, ribs ramify in two branches. The ramification point is located, as a rule, in the center of the lateral side or, less frequently, on the bend from the dorsal to lateral side. On the dorsal side, ribs slightly bending backward become lower and disappear at the beginning of the trunk.

Size (mm) and angle (degree) parameters (see Fig. 5 for explanations).

Specimen no.	H	ht*	ht	wt	hb	wb	hh	wh	α_1	α_3	δ
1/96	28	2	4.5		5.5		6		15	12	70
2/96			5.2	5.1	6	7	6.5	6.6	15	13	70
2478/3390 Holotype	<23	2	4.3		5		5.5		15	12	65
2480/3390			4.5	4	5.3		6		13	12	70
3/96			4.3		5				15	12	70
4/96					5	6			15	13	

The lobe line is characterized in the genus description.

Comparison. Similar specimens are pictured in work by Koenen under the name *Ancyloceras biplicatum* (Koenen, 1902: p. 379; Plate XLI, figs. 2a–b, 8a–b). Klinger and Kennedy (1977) attributed the *biplicatum* form to the genus *Toxoceratoides*. Aguirre Urreta (1986) arbitrarily assigned it to the same genus, particularly emphasizing this fact (p. 296). Unfortunately, samples collected by Koenen have not been preserved.

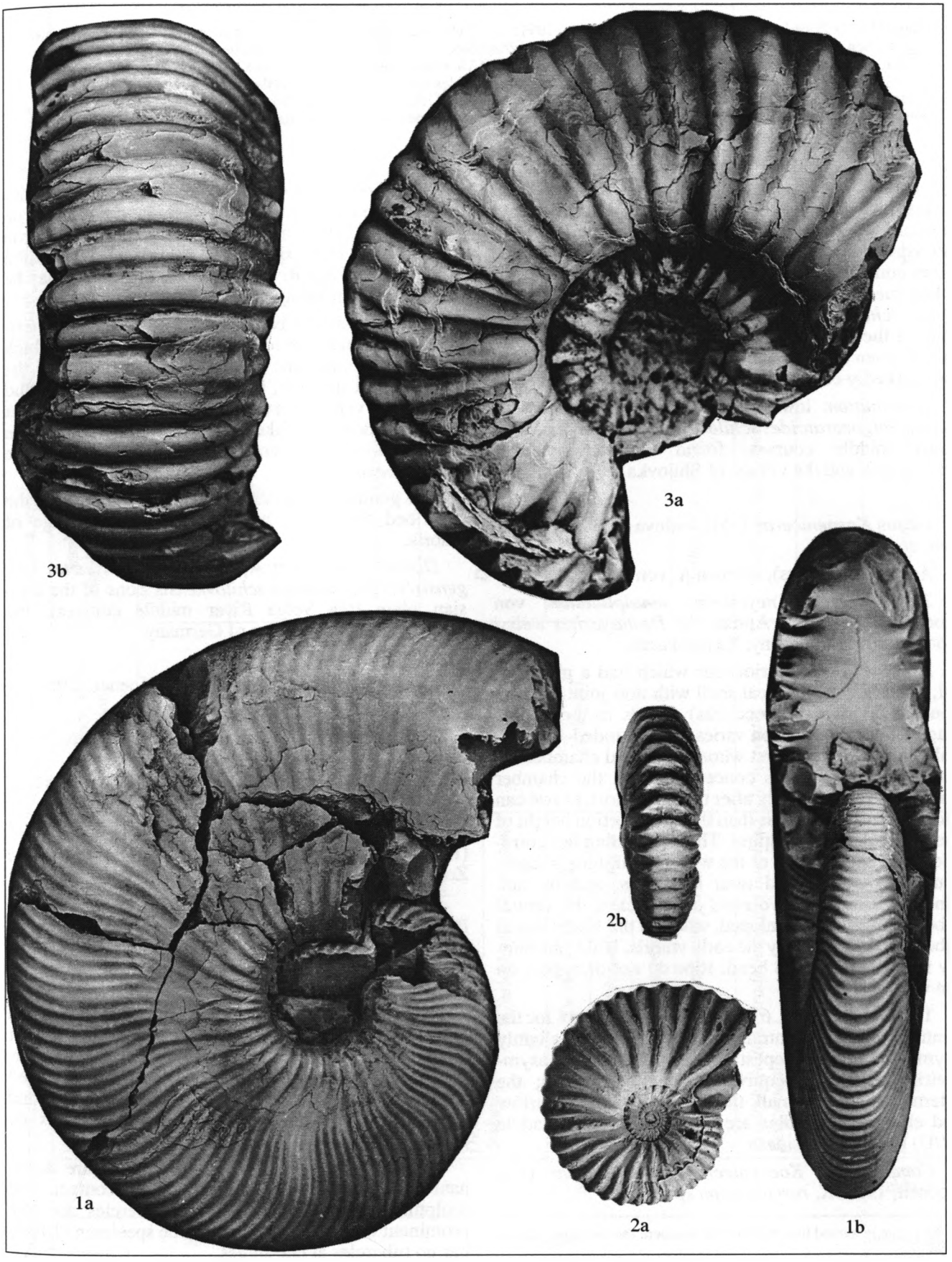
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Plate IV. (1) *Paradeshayesites similis* (Bogdanova, 1991), sp. 23/96 MPG MGU, side (a) and aperture (b) views, Ul'yanovsk region, Sengilei, lower Aptian, the *Deshayesites volgensis/Ancyloceras matheronianum* Zone (collected by Uspenskii and Shumilkin). (2) *Deshayesites saxbyi* Casey, 1964, sp. A/3390 PIN RAS, side (a) and ventral (b) views, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by K.A. Kabanov). (3) *Cheloniceras* ex. gr. *cornuelianum* (d'Orbigny, 1841), sp. 27/96 MPG MGU, side (a) and ventral (b) views, Ul'yanovsk region, Sengilei, lower Aptian, the *Deshayesites deshayesi/Audouliceras renauxianum* Zone (collected by Uspenskii and Shumilkin).

On images, they exemplify only the hamulus fragments, thus giving only a rough idea about the *Volgoceratoides biplicatum* morphology. It is evident from description that the latter has only one row of tubercles. From comparison with our, more complete specimens, it becomes clear that lateral tubercles of the *Volgoceratoides schilovkensis* disappear long before the beginning of the hamulus. Moreover, specimens studied by von Koenen have frontal branches of bifid ribs, which are markedly curved forward on the hamulus.

Distribution: lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone of the Volga River middle courses, found in sections near Ul'yanovsk and the village of Shilovka.

Genus *Koeneniceras* I. Michailova et Baraboshkin, gen. nov.[3]

Ancyloceras (pars), Koenen A. von 1902, p. 331.

Type species: *Ancyloceras tenuiplicatum*: von Koenen, 1902; lower Aptian, the *Deshayesites weissi* Zone, northern Germany, Kastendamm.

Diagnosis. Small criocones which had a planospiral, slightly asymmetrical shell with non-joint (macroconchs) or joint (microconchs) whorls at the middle stages. The cross section varies from rounded–hexagonal to nearly round. First whorl and initial chamber are unknown, though it is conceivable that the chamber could form immediately after the first whorl, as one can judge from a small (less than 0.5 mm) section height of the initial part of the spiral. The living chamber corresponds to about a half of the whorl. Sculpture is represented by simple and rarer bifid ribs, and by rare pinches. Ribs have two rows of tubercles: the ventral tubercles are well developed, whereas the lower lateral tubercles decorate only the early whorls. Bifid ribs ramify near the umbilical bend. Ribs do not disappear on the ventral side.

The lobe line (Fig. 6) has trifid lobes, except for the ventral ones. The ventral lobe (V) is shallow, slightly asymmetrical; the deepest umbilical lobe (U) is asymmetrical, with non-equivalent lateral denticles; the internal lobe (I) is small; the dorsal (D) lobe is narrow and elongated. Saddles are bifid, the external saddle (V/U) has unequal sides.

Composition: *Koeneniceras tenuiplicatum* (von Koenen, 1902, *K. rareplicatum* sp. nov.

[3] The genus is named honor of A. von Koenen, the German paleontologist.

Comparison and remarks. As we know, nobody has ever referred so far to the species *tenuiplicatum* von Koenen, 1902. Judging from fragments pictured by von Koenen and from specimens of better preservation found in the Volga River basin area, this genus may be thought of as differing from all known genera.

The most similar is the Hauterivian genus *Aegocrioceras*, the endemic form of northern Europe, which has larger dimensions and shows a lesser rate of the whorl height increase, more rarely arranged ribs, and presence of ventrolateral tubercles only and their earlier disappearance. It is likely that representatives of the *Aegocrioceras* genus could be ancestral forms for the genus *Koeneniceras*.

The genus *Hemihoplites* also slightly resembles the described taxon, but it exhibits a faster linkage of whorls.

Distribution: lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone of the Russian plate (the Volga River middle courses), the *Deshayesites weissi* Zone of Germany.

Koeneniceras tenuiplicatum (von Koenen, 1902)

Plate VI, figs. 2–5

Ancyloceras ? cf. *brevispina*: von Koenen, 1902, p. 365, Plate XL, fig. 4

Ancyloceras tenuiplicatum: von Koenen, 1902, p. 377, Plate XLV, fig. 11; Plate LIII, fig. 5

Lectotype: specimen pictured in Plate VIII, fig. 5 (Koenen, 1902), lower Aptian, the *Deshayesites weissi* Zone, northern Germany, Kastendamm.

Material: eight almost whole specimens of good preservation and one large imprint.

Description. Shells are small, slightly asymmetrical, of the crioceratid morphology (up to 40–50 mm in diameter). The cross section is rounded–hexagonal, slightly elongated in height and flattened at the venter. Ribs are frequent, undivided or rarely (up to 1–5 per whorl) bifid, crossing the ventral side with some lowering. At the early whorls, there are two pairs of flattened tubercles: the ventrolateral disappearing by ageing and ventral. Pinches are rare. Micro- and macroconchs are distinguishable.

Macroconchs of the largest dimensions are dominant and abundant. Whorls are out of contact, and sculpture appears relatively late. Tubercles are less prominent than on microconchs. One specimen (10/96) has no tubercles at late stages.

Table 2. Ammonoid zonations and correlation between lower Aptian deposits of the Ul'yanovsk region (the Volga River basin), England, and Germany

Stage	Substage	Casey, 1961; Casey *et al.*, 1998 England		Kemper, 1995 Germany	Baraboshkin, 1998, Baraboshkin *et al.*, 1999 Ul'yanovsk	This work Ul'yanovsk region	
		Zone	Subzone	Zone	Zone	Zone	
						Deshayesitidae	Ancyloceratidae
Aptian	Middle (part)	*Cheloniceras martinoides* (part)	*Epicheloniceras debile*	*Tropaeum drewi* + *Tropaeum tenuinodosum*	*Aconeceras nisum*	*Aconeceras nisum*	
	Lower	*Tropaeum (Tropaeum) bowerbanki*	*Cheloniceras meyendorffi*	*Tropaeum bowerbanki* + *Dufrenoyia furcata*	*Tropaeum bowerbanki*		*Tropaeum bowerbanki*
			Dufrenoyia transitoria				
		Deshayesites deshayesi	*Deshayesites grandis*	*Deshayesites deshayesi* ?	*Deshayesites grandis*	*Deshayesites deshayesi*	*Audouliceras renauxianum*
			Cheloniceras parinodum				*Proaustraliceras tuberculatum*
		Deshayesites forbei	*Deshayesites annelidus*		Beds with *Deshayesites consobrinoides*	*Deshayesites volgensis*	*Ancyloceras matheronianum*
			Deshayesites callidiscus				
			Deshayesites kiliani				
			Deshayesites fittoni		*Deshayesites deshayesi*		*Volgoceratoides schiloivkensis*
		Prodeshayesites fissicostatus	*Prodeshayesites obsoletus*	*Deshayesites tenuicostatus*	*Deshayesites forbesi*	*Deshayesites tenuicostatus*	
			Prodeshayesites bodei	*Deshayesites bodei*	?	?	
Barremian	Upper	Veld		*Parancyloceras bidentatum*	*Oxyteuthis lahuseni*	*Oxyteuthis lahuseni*	

Microconchs exhibit a shortened stage with the poorly developed sculpture. Whorls stop to get into contact when D = 20 mm. Ribs are coarser and have tubercles more developed than those of macroconchs.

Dimensions (mm) and angles (degree)

Specimen no.	D	Ud	H	sbw	W	α_1	α_3
2474/3390	29	12.7	11.4	1	12	–10	5
2297/3390	35	14	12	1.5		–5	
5/96	31	13	11.5	1	8	–10	–5
6/96	~19	9	6		~6	–15	
7/96	13	8	4.4		4	–5	
8/96	20	9	9		8.5	–10	–5
9/96	20.5	8	7		8	–15	
11/96	24.5	11.5	9		7	–5	5
10/96	27.7	11	10.5	2	8	–10	–5

The lobe line is characterized in the genus description.

Comparison. In distinction from *Koeneniceras rareplicatum* sp. nov., ribbing of the described species is more frequent.

Remarks. Among specimens of *"Ancyloceras tenuiplicatum,"* pictured by Koenen (1902, p. 377, Plate XLV, fig. 11; Plate LIII, fig. 5), specimen no. 5 in Plate LIII is proposed to be the lectotype because specimen fig. 11 in Plate XLV likely had some defect in growth. Unfortunately, we failed to find type material of von Koenen in the collection of the Goettingen University Museum.

Distribution is similar to that of the genus *Koeneniceras.*

Koeneniceras rareplicatum I. Michailova et Baraboshkin, sp. nov.[4]

Plate VI, fig. 12

Ancyloceras cf. *tenuiplicatum*: von Koenen, 1902, p. 379, Plate LIII, fig. 4.

Holotype: 33/96 MPG MGU, the Volga River right bank, the Ul'yanovsk region, the village of Shilovka; lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone.

Material: one whole well-preserved specimen (microconch).

Description. Shell is small, evolute, and slightly asymmetrical. The cross section is rounded–hexagonal. Ribs are single, embossed, and widely spaced; a bifid rib may occur on the whorl. All ribs intersect the ventral side with a noticeable lowering. Two pairs of small tubercles (ventrolateral and ventral) are present on the early whorls and disappear by ageing.

[4] Name is from Latin rarus (rare) and plicatilis (folded).

By analogy with the type species, the specimen from our collection should be attributed to microconchs: its whorls are in contact and ribbing is rough.

Dimensions (mm) and angles (degree).

Specimen no.	D	Ud	h	w	α_1	α_3
33/96 Holotype	16	7	5.5	6	–10	5

Comparison. Main points are above when describing the species *K. tenuiplicatum* von Koenen. In distinction from new *tenuiplicatum* form, species *"Ancyloceras"* cf. *tenuiplicatum* (Koenen, 1902, Plate LIII, no. 4) has whorls slowly increasing in height and rare ribs. By analogy with the type species, it can be regarded as macroconch of the new species.

Distribution is similar to that of the genus *Koeneniceras.*

CONCLUSION

Sections of lower Aptian deposits in the Volga River middle courses are unique in their completeness and paleontological characteristics. The comprehensive study of these deposits and layer-by-layer collection of ammonites made it possible to elaborate the new biostratigraphic scheme for lower Aptian deposits in the Russian plate. The scheme is as detailed as its analogs elaborated in Turkmenistan, Great Britain, and Germany. In contrast to the last regions, the scheme proposed for the Volga River middle courses is based on evolutionary trends of two parallel ammonite groups: monomorphic neritic Deshayesitidae and heteromorphic subplanktonic, probably benthopelagic *Ancyloceratidae.* This allows correlation between stratigraphic scales of shallow-water and relatively deep, epipelagic sediments accumulated in the basin.

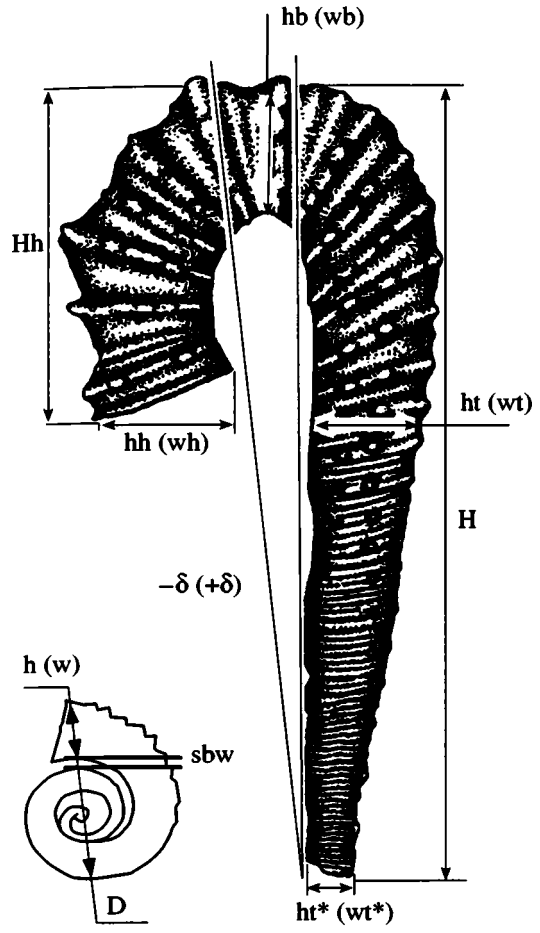


Fig. 5. Morphometric parameters of ammonites: (H) shell height; (Hh) hamulus height; (ht) height of the trunk cross section; (wt) width of the trunk cross section; (ht*) height of cross section at the beginning of trunk; (wt*)width of cross section at the beginning of the trunk; (hb) height of cross section at the bend of hamulus; (wb) width of cross section at the bend of hamulus; (hh) heights of cross section at the end of hamulus; (wh) width of cross section at the end of hamulus; (δ) angle of hamulus bending outward (+) or toward (–) the trunk; (D) diameter of the spiral part; (Ud) umbilicus diameter; (h) height of whorl at the end of spiral; (w) width of whorl at the end of spiral; (sbw) spacing between whorls at the end of spiral.

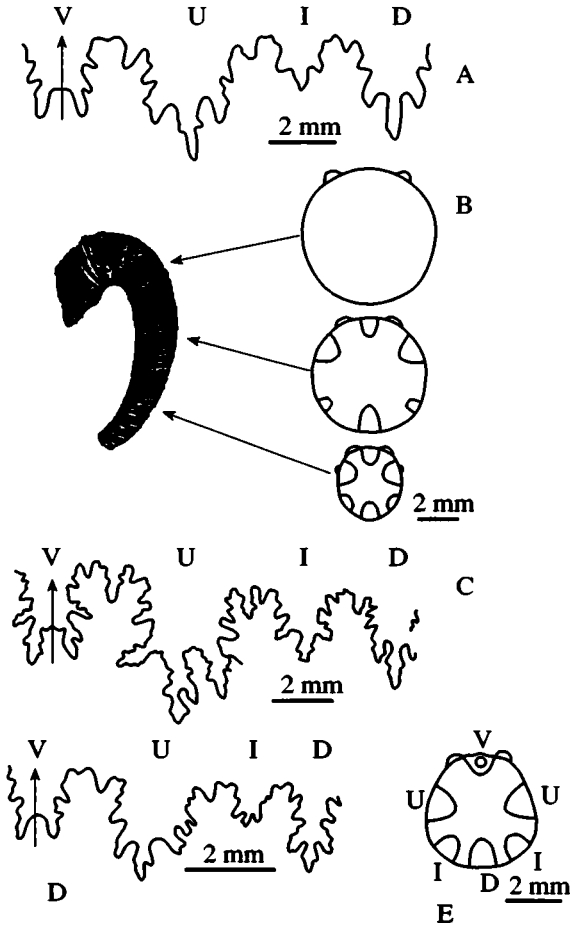


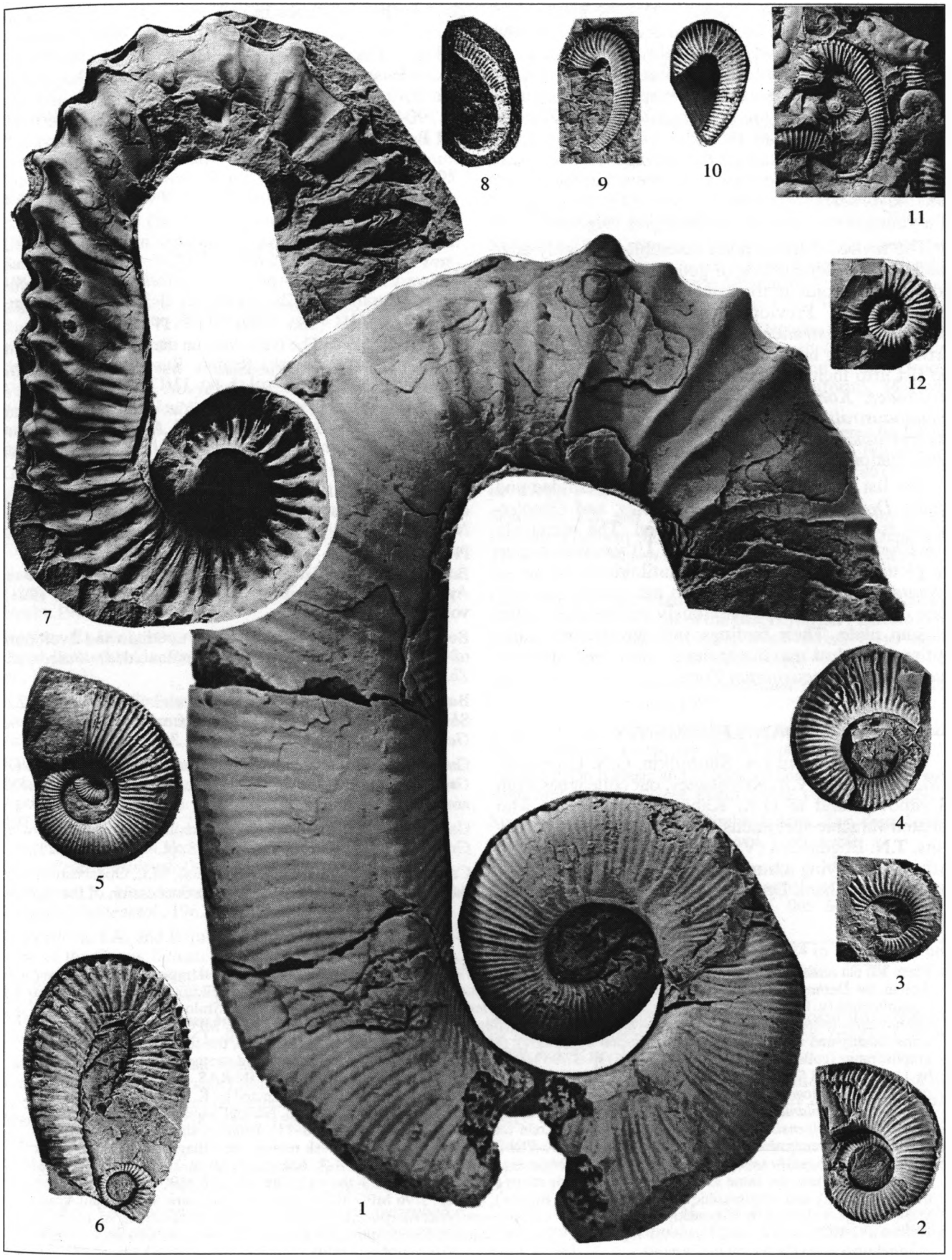
Fig. 6. Lobe lines and cross section configuration of studied ammonites. (A, B) *Volgoceratoides schilovkensis* I. Michailova et Baraboshkin, sp. nov., specimen 2/96 MPG MGU, lower Aptian, the *Deshayesites forbesi/Volgoceratoides schilovkensis* Zone, Ul'yanovsk region, the village of Shilovka: (A) lobe line when h = 4.7 mm and w = 4.4 mm, (B) changes of cross section shape when h = 3.1, h = 4.9, and h = 5.7 mm respectively; (C–E) *Koeneniceras tenuiplicatum* (von Koenen, 1902), the same locality and age: (C) specimen 2474/3390 PIN RAS, lobe line when h = 7.5 mm, (D) specimen 6/96 MPG MGU, lobe line when h = 3.9 mm and w = 3.6 mm, (E) the same specimen, cross section when h = 3.8 mm and w = 3.3 mm.

The *Deshayesites tenuicostatus* Zone was distinguished for the first time in the study region. Its ammonite assemblage is similar to that from the synonymous zone of northern Germany that radically changes our understanding of the sea transgression direction that caused fauna migration at the beginning of the Aptian.

The lower Aptian deposits of the study region were previously attributed to the *Deshayesites deshayesi*

Plate V. (1) *Proaustraliceras tuberculatum* (Sinzow, 1872), sp. 2563/3390 PIN RAS, side view, magnification ×0.7, Ul'yanovsk, lower Aptian, the *Deshayesites deshayesi/Proaustraliceras tuberculatum* Zone (collected by K.A. Kabanov). (2) *Lithancylus russiensis* I. Michailova et Baraboshkin, 2001, sp. 15/96 MPG MGU, holotype, side view, magnification ×0.7, Ul'yanovsk region, Sengilei, lower Aptian, the *Deshayesites volgensis/Ancyloceras matheronianum* Zone (collected by Uspenskii and Shumilkin). (3) *Pseudoaustraliceras pavlovi* (Wassiliewski, 1908), sp. 2506/3390 PIN RAS, ventral (a) and side (b) views, Ul'yanovsk, Solov'ev Ovrag, lower Aptian, the *Deshayesites volgensis/Ancyloceras matheronianum* (collected by K.A. Kabanov). (4) *Sinzovia trautscholdi* (Sinzow, 1870), sp. 1665/3390 PIN RAS, side (a) and aperture (b) views, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by K.A. Kabanov).

4b
4a
3b
1
3a
2


8
9
10
11
12
7
5
4
3
6
1
2

Zone, but we may state at present that the greater part of the section, the shale horizon included, yields the diverse ammonite assemblage of the *Deshayesites volgensis* Zone. The species composition of deshayesitids from this zone is similar in many respects to the ammonite assemblage from the *Deshayesites forbesi* Zone of Great Britain. Moreover, the index *forbesi* species from Great Britain is very similar to the *volgensis* species. If the identity of these forms can be proved in the future, then the *forbesi* form will be transferred to the category of a younger synonym of *Deshayesites volgensis*.

The revised Cephalopoda assemblages highly augmented our understanding of generic and species diversity of ammonoids in the lower Aptian deposits of the Russian plate. Previously, only *Ancyloceras, Tropaeum*, and *Australiceras* genera were known in this territory among heteromorphic ammonites. At present, the list also includes representatives of genera *Volgoceratoides, Koeneniceras, Ancyloceras, Lithancylus, Pseudoaustraliceras, Proaustraliceras, Audouliceras, Toxoceratoides, Tropaeum*, and a series of new forms investigation of which is now in progress.

The list of forms representing Deshayesitidae and genera *Deshayesites, Paradeshayesites*, and *Obsoleticeras* is also substantially extended. The extremely rare *Cheloniceras* forms from the Ul'yanovsk region are pictured for the first time. Nautiloids of the genus *Cymatoceras* found in the region are unique, because none of them have been previously encountered in the Russian plate. Their findings indicate that the early Aptian basin had maximum depth in the period of the *Audouliceras renauxianum* Zone.

ACKNOWLEDGMENTS

We are grateful to I.A. Shumilkin, G.N. Uspenskii, V.M. Efimov, and V.A. Krivosheev, our colleagues from Ul'yanovsk, and to G.K. Kabanov (PIN RAS) who donated valuable specimens from their personal collections. T.N. Bogdanova (VSEGEI) fruitfully cooperated with us in studying ammonites of the Volga River basin area. We also thank Dr. H. Jahnke (Goettingen University) for his assistance in work on collection of A. von Koenen. Cephalopods are pictured by V.A. Antonova and A.V. Mazin (PIN RAS). The work was supported by the Russian Foundation for Basic Research, project nos. 00-05-64738, 01-05-64641, 01-05-64642, and 02-05-79040, as well as by DAAD (ref. 325) and the Federal Program "Integration."

Reviewers I.A. Basov and K.I. Kuznetsova

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Plate VI. (1) *Audouliceras renauxianum* (d'Orbigny, 1842), sp. 31/96 MPG MGU, side view, Ul'yanovsk region, Sengilei, lower Aptian, the *Deshayesites deshayesi/Audouliceras renauxianum* Zone (collected by Uspenskii and Shumilkin). (2–5) *Koeneniceras tenuiplicatum* (v. Koenen, 1902): (2) sp. 5/96 MPG MGU, side view, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by G.K. Kabanov); (3) sp. 6/96 MPG MGU, side view, the same locality and stratigraphic range (collected by Efimov); (4) sp. 2474/3390 PIN RAS, side view, the same locality and stratigraphic range (collected by K.A. Kabanov); (5) sp. 37/96 MPG MGU, side view, the same locality and stratigraphic range (collected by Uspenskii and Shumilkin). (6) *Toxoceratoides royerianus* (d'Orbigny, 1842), sp. 2510/3390 PIN RAS, side view, Ul'yanovsk region, Sengilei, lower Aptian, the *Deshayesites deshayesi/Audouliceras renauxianum* Zone (collected by K.A. Kabanov). (7) *Ancyloceras matheronianum* d'Orbigny, 1842, sp. 12/96 MPG MGU, side view, Ul'yanovsk region, Novoul'yanovsk, lower Aptian, the *Deshayesites volgensis/Ancyloceras matheronianum* Zone (collected by Krivosheev). (8–11) *Volgoceratoides schilovkensis* I. Michailova et Baraboshkin, sp. nov.: (8) sp. 2481/3390 PIN RAS, side view, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by K.A. Kabanov); (9) sp. 2478/3390 PIN RAS, holotype, side view, the same locality and stratigraphic range (collected by K.A. Kabanov); (10) sp. 2/96 MPG MGU, side view, the same locality and stratigraphic range (collected by Efimov); (11) sp. 1/96 MPG MGU, side view, the same locality and stratigraphic range (collected by Uspenskii and Shumilkin). (12) *Koeneniceras rareplicatum* I. Michailova et Baraboshkin, sp. nov., sp. 33/96 MPG MGU, side view, Ul'yanovsk region, the village of Shilovka, lower Aptian, the *Deshayesites volgensis/Volgoceratoides schilovkensis* Zone (collected by Efimov).

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Phytoplankton Provinces of the Tremadocian Time

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Received October 26, 2000

Key words: acritarchs, phytoplankton province, southern Urals, Tremadocian, Alimbet Horizon, Kolnabuk Formation, Kuragan Formation, Terekla Formation.

Analysis of acritarch assemblages from coeval Upper Cambrian and Lower Ordovician beds of different regions of the planet revealed their different composition. Specialists who studied the assemblages explain this difference by existence of several phytoplankton provinces, where basins differed in the water temperature and other parameters.

Volkova (1995, 1997) summarized pioneering publications of the 1970s–1980s and supplemented them by her own data on the Baltic region, the Leningrad, Yaroslavl', and some other areas. As a result, three Late Cambrian–Tremadocian phytoplankton provinces have been distinguished (Volkova, 1997): the cool-water Mediterranean (Peri-Gondwanan), moderate Baltic, and warm-water Equatorial.

Volkova (1997) marked the Late Cambrian and Tremadocian acritarch localities of these provinces on the palinspastic map that had been compiled by Erdtmann for the Early Ordovician (Fig. 1). The map shows that the localities of the Baltic province are concentrated in the northeastern Europe (the Arkhangel'sk and Yaroslavl' areas) but not farther than 40°–41° E (in modern coordinates, Fig. 1). The Lower Ordovician deposits have been long known in the southern Urals. Despite their faunal remains (mostly trilobites and brachiopods), they received ambiguous and even contradictory stratigraphic interpretations. This is especially true for deposits, which have been considered as Cambrian in age.

The deposits under consideration are confined to complicatedly dislocated Sakmara nappe and its southern, Kosistek–Aksu continuation to the south of the latitudinal course of the Ural River (the Kazakhstan part of the Urals). In this area, there are type sections of these deposits, including those, where the Urals regional horizons were established (*Stratigraficheskie...*, 1993).

We obtained palynological data from some main sections of the area (Fig. 2), which essentially corrected the Cambrian–Lower Ordovician stratigraphy of the Urals (Chibrikova and Olli, 1996a, 1996b, 1997). These data also supplement the available information on the Tremadocian phytoplankton provinces. According to acritarch distribution in the Lower Ordovician hypostratotype at the Alimbet River (the left bank of the Ural River, Fig. 2), it appeared to be more expedient to establish the Alimbet Horizon (Tremadocian) instead of the Kidryas and Kolnabuk horizons and the Akbulaksai Horizon (lower Arenigian?) instead of the Kuagach Horizon (Chibrikova and Olli, 1996a, 1996b).

The Alimbet Horizon stratotype located at the eponymous river contains acritarchs of early and late Tremadocian age, whereas some other sections can be

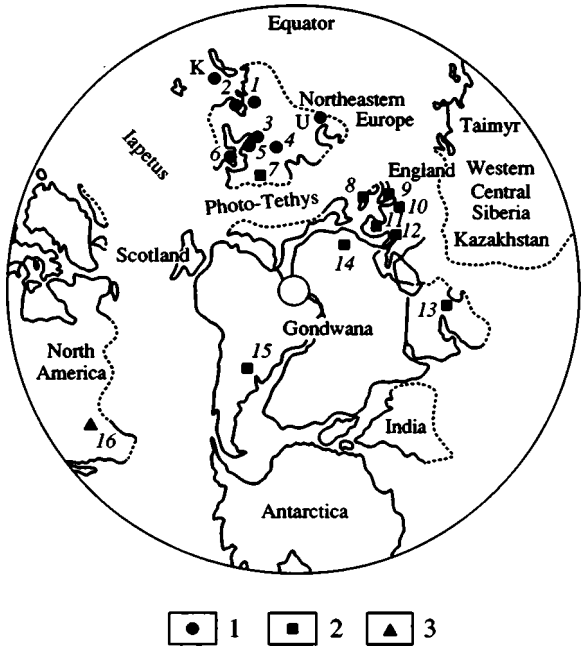
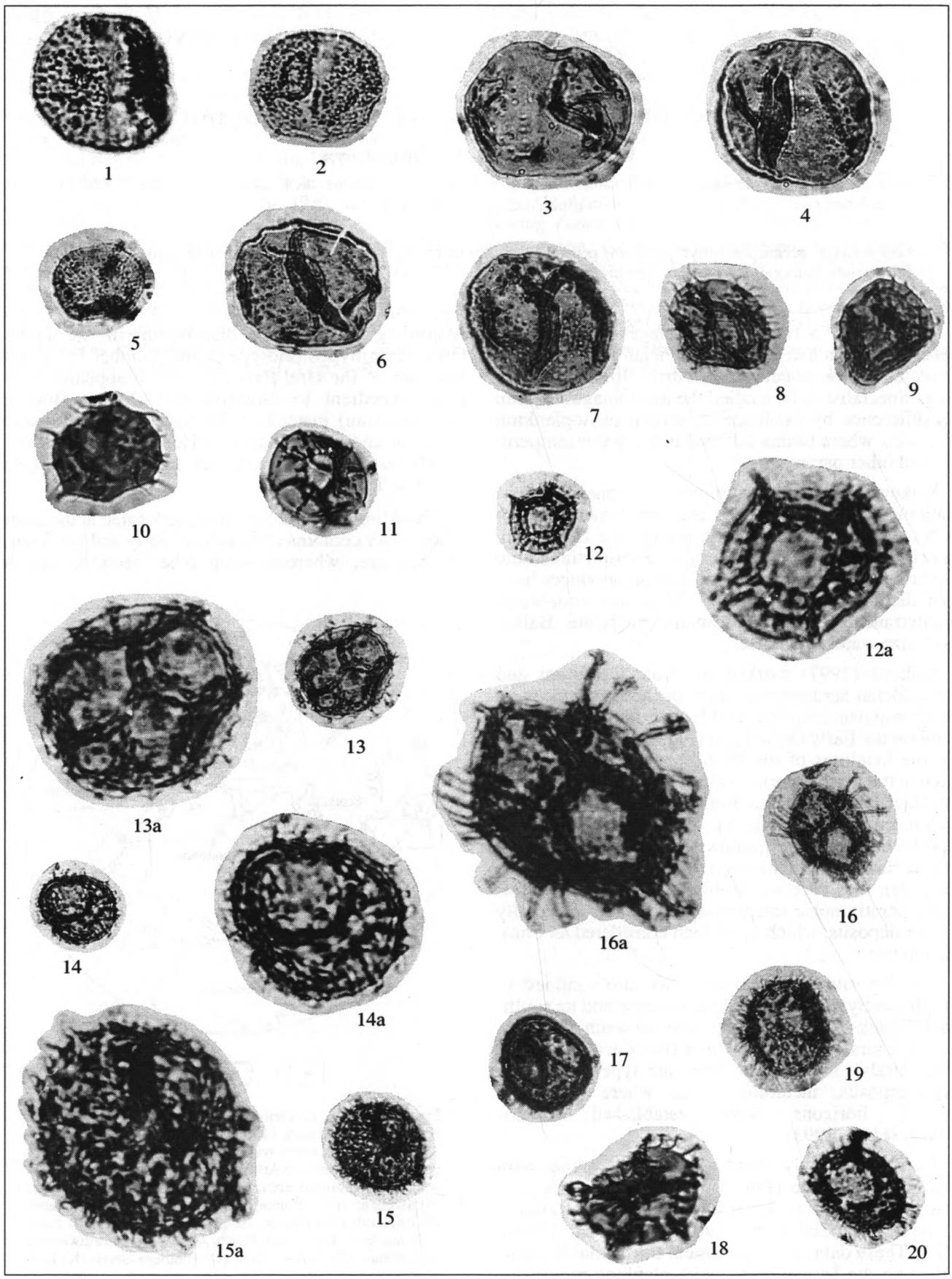


Fig. 1. Localities of Tremadocian acritarchs characterizing, according to Volkova (1997), the Baltic (1), Peri-Gondwanan (2), and warm-water (3) provinces. Localities by geographic names: (*1*) Arkhangel'sk area; (*2*) northern Norway; (*3*) Leningrad area; (*4*) Yaroslavl area; (*5*) Estonia; (*6*) Sweden; (*7*) Poland; (*8*) eastern Newfoundland; (*9*) England; (*10*) France, Belgium; (*11*) Spain; (*12*) Italy; (*13*) northern Iran; (*14*) North Africa; (*15*) northwestern Argentina; (*16*) Texas) USA; (U) southern Urals; (K) Kolguev Island.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assigned, based on acritarchs, to the undifferentiated Tremadocian. The lower part of the Alimbet Horizon (the Alimbet Farm section, Fig. 2) yields the Early Ordovician acritarch assemblage with dominant *Acanthodiacrodium angustum* (Downie), *A. comptulum* Rasul, *A. papillare* Tim., *Acanthoritidodiacrodium unigeminum* Tim., etc. (plate).

The presence of *Acanthodiacrodium angustum* (Downie) that associates with conodont species *Cordilodus proavus* Müll. in some sections (Volkova, 1995, 1997, 1999) is of a special significance. Both forms are index species of zones, the basal levels of which mark the lower boundary of the Tremadocian Stage. The lower part of the Alimbet Horizon was formerly regarded as barren of faunal remains, and it was attributed to the lower Tremadocian only when we found acritarchs at this level.

The upper part of the Alimbet Horizon yields the late Tremadocian acritarchs, such as *Cymatiogalea multiclaustra* Deunff, *C. cuvillieri* Deunff, *Priscogalea plenae* (Vand.), *P. serrata* (Naum.), and *P. spectatissima* (N. Umn.), which have large polar openings or pylomes (plate). Korinevskii (1989) published the list of faunal forms found here by him and other researchers. These were mostly trilobites *Asaphellus alimbeticus* Bal., *Promegalaspides kazachstanensis* Bal., *Ceratopyge forficula* Cars, *C.* sp., *Schumardia pusilla* Cars, *Sch. oelandica* Möb., *Leiagnostus alimbeticus* Bal., *Gerragnostus sidenblandhi* (Linn.), *Parabolinella limitis* Brög., *Apatokephalus serratus* (Boeck), and *Cyclognatus mycropygus* Linn., brachiopods *Altorthis kinderlensis* Andr., *Paldiskia* sp., *Siphonotreta uralenis* Lerm., and *Euritreta (?) chabakovi* (Lerm.), and graptolites *Dictyonema uralensis* Obut. In spite of numerous endemics, the assemblage is similar to the late Tremadocian *Ceratopyge* fauna of Scandinavia (Varganov *et al.*, 1973; Korinevskii, 1989).

The middle part of the Alimbet Horizon of the same section contains a mixed assemblage of acritarchs. These are envelopes with big pylomes and diacrodian forms, which associate with forms of a greater Lower Ordovician range (Chibrikova and Olli, 1996a, 1996b).

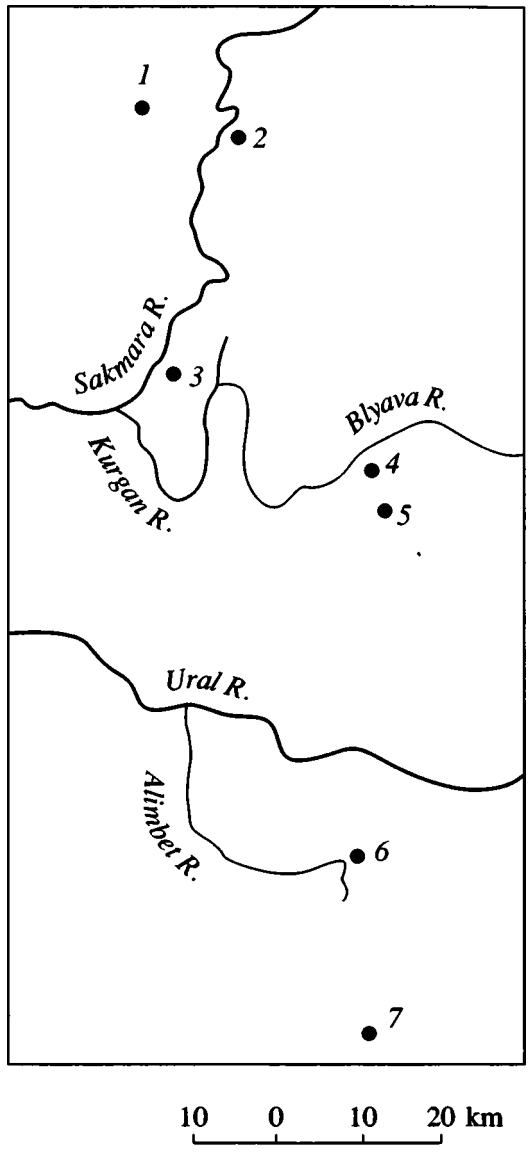

Fig. 2. Localities of Tremadocian acritarchs in the southern Urals (Bashkortostan, Orenburg area, and Kazakhstan): (*1*) Kisi-Yara Creek, the tributary of the Vazyam River upstream of the Bashkir Bazyam Village; (*2*) left bank of the Sakmara River opposite to the Bashkir Vazyam Village; (*3*) Mt. Biktash; (*4*) Blyava River; (*5*) Shiyakla Creek; (*6*) Alimbet Farm at the eponymous river; (*7*) Krasnyi Ravine.

Acritarchs from the Tremadocian Alimbet Horizon stratotype, the western slope of the southern Urals (magnification ×600, except for that specially indicated).

(1, 2) *Acanthodiacrodium angustum* (Downie) Combaz: (1) specimen IG USC RAS, no. 5673d, (2) specimen IG USC RAS, no. 5673a; (3, 4) *Acanthodiacrodium comptulum* Rasul: (3) specimen IG USC RAS, no. 5673d, (4) specimen IG USC RAS, no. 5673b; (5) *Acanthodiacrodium papillare* Tim, specimen IG USC RAS, no. 5673d; (6, 7) *Acanthoritidodiacrodium unigeminum* Tim., specimen IG USC RAS, no. 5673b; (8, 9) *Acanthodiacrodium* cf. *costatum* Burm.: (8) specimen IG USC RAS, no. 5710b, (9) specimen IG USC RAS, no. 5710a; (10) *Arbusculidium* sp., specimen IG USC RAS, no. 5710a; (11–13a) *Cymatiogalea cuvillieri* Deunff: (11) specimen IG USC RAS, no. 5710a, (12–13a) specimen IG USC RAS, no. 6248a, (12a and 13a) ×1500, (13a) side view; (14–15a) *Priscogalea plenae* (Vand.) Arist.: (14, 14a) specimen IG USC RAS, no. 5670a, (14a) ×1500, (15, 15a) specimen IG USC RAS, no. 5670b, side view, (15a) ×1500; (16, 16a) *Cymatiogalea multiclaustra* Deunff, specimen IG USC RAS, no. 6248a, (16a) ×1500; (17) *Priscogalea serrata* (Naum.) Arist., specimen IG USC RAS, no. 5710a; (18) *Cymatiogalea* cf. *multiclaustra* Deunff, specimen IG USC RAS, no. 5710a; (19, 20) *Priscogalea spectatissima* (N. Umn.) Olli in litt.: (19) specimen IG USC RAS, no. 5670b, (20) specimen IG USC RAS, no. 6616a.

Late Tremadocian acritarchs in association with other coeval fossils were also found in the "Kolnabuk Horizon" stratotype at the Blyava River and to the south, at the Shiyakla Creek (Fig. 2). We consider their host deposits as corresponding to the Kolnabuk Formation that spans the upper part of the Alimbet Horizon (Chibrikova and Olli, 1996a, 1996b). The Kolnabuk Formation stratotype contains abundant *Priscogalea serrata* (Naum.), *P. spectatissima* (N. Umn.), and *P. grumosa* Arist. in association with single *Cymatiogalea* Deunff and *Stelliferidium* Deunff (Chibrikova and Olli, 1996a). Coexisting trilobites represent species *Ceratopyge* sp., *Schumardia oelandica* Möb., *Gerragnostus sidenblandhi* (Linn.), *G.* sp., *Protopliomerops speciosus* (Dalm.), *Euloma* cf. *kelleri* Bal., and *Apatokephalus* sp. (Varganov *et al.*, 1973).

At the Vazyam River (the right tributary of the Sakmara River) and at the left bank of the Sakmara River (opposite to the Bashkir Vazyam Village, Fig. 2), the lower part of the sequence that has been previously considered as the upper Arenigian–Llanvirnian Kuragan Formation barren of fossils (*Stratigraficheskie...*, 1993), was assigned to the Kolnabuk Formation based on the found acritarchs (Chibrikova and Olli, 1997). The acritarchs found date the deposits of the Krasnyi Ravine, the left bank of the Ural River in the Kazakhstan part of the southern Urals, and those at the Mt. Biktash, the left bank of the Sakmara River (Fig. 2), back to the undifferentiated Tremadocian. Earlier these deposits with Tremadocian acritarchs were incorrectly united with archaeocyathid limestones into the Lower Cambrian Terekla Formation (Klyuzhina, 1990; *Stratigraficheskie...*, 1993).

According to data published by Volkova (1995, 1997, 1999) and other authors (Aristova, 1980; Verichev *et al.*, 1990; Umnova, 1973), the acritarch assemblages under consideration are similar to those from the Lower Ordovician deposits of the Baltic region and central Russian platform. This suggests that the southern Urals represented a part of the moderate Baltic phytoplankton province in the Tremadocian time. Thus, this province extended at least to 57° E (the southern Urals, Fig. 1, Site U), but not to 40°–41° E, as previously estimated.

However, we do not assume existence of a single sedimentation basin over the Russian platform and the Urals. There might be two separate basins similar in the water temperature and sedimentation environments.

As mentioned above, the Tremadocian sections of the southern Urals are confined to the nappes thrust over younger deposits from the east. These nappes, particularly the Sakmara nappe, are thought to have been rooted in the Main Uralian fracture zone (Puchkov, 2000) behind the Ural-Tau Ridge, i.e., 30–35 km to the east of their present-day position. Therefore, we can suggest that the Baltic phytoplankton province extended eastward for the same distance. There are no records of Tremadocian acritarchs in the Trans-Urals and Siberia.

Acritarch assemblages of similar composition, which are known from the Tremadocian deposits of the Kolguev Island (Rudavskaya and Raevskaya, 1994), indicate that the last area also belonged to the Baltic phytoplankton province (Fig. 1, Site K). Thus, in northern European Russia, this province reached approximately 49° E (in the modern coordinates).

Reviewers N.A. Volkova and B.S. Sokolov

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